

Geology of Idaho



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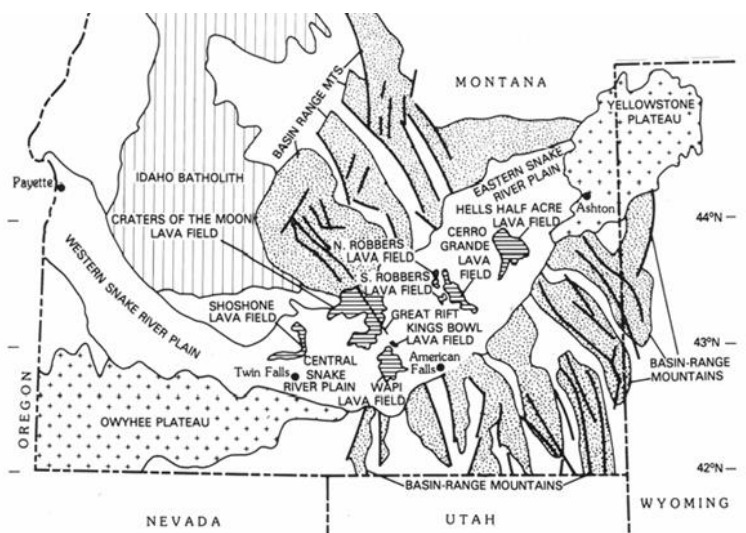
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Basic Geology with Idaho Examples

Introduction to Idaho Geology

Geology is the science which deals with the origin, structure and history of the Earth. This also includes the study of past life as recorded in the rocks. In geologically recent times, Idaho has experienced major earthquakes, catastrophic floods, the Pleistocene Ice Age and volcanic eruptions. Idaho is one of the few states with excellent examples of almost every type of geologic landform, structure, rock type and mineral deposit. Idaho's oldest rocks are more than 2500 million years old and contain a record of all the events that transpired during that vast period of geologic time.

Southern Idaho contains a remarkably diverse group of geologically significant features and landscapes—perhaps more than any comparably sized area in the world. The regional geologic setting includes world-class tectonic features such as the basin- and-range faults, the Idaho- Wyoming Thrust Belt, the track of the Yellowstone hotspot, the Snake River Plain and the Idaho Batholith. Rocks in the area range in age from 2200- year-old volcanic rocks at Craters of the Moon National



Generalized geologic map of southern Idaho showing major geologic and physiographic features.

Monument to the 2.5 billion-year-old gneiss in the Albion Range. The most recent geologic event to affect the area is the catastrophic Bonneville flood that swept across southern Idaho 14,400 years ago and left spectacular large-scale erosional and depositional features along the Snake River. This unique concentration of so many diverse and scientifically significant features such as the Big Wood River, the Snake River Canyon, the Gooding City of Rocks, the Bonneville Flood, Black Butte Shield Volcano, Shoshone Caves, Quaternary lava fields and the volcanic rift system has created a remarkable outdoor geological museum.

The Earth is approximately 4.6 billion years old, originating about the same time as the other planets in our solar system. The oldest rocks found on Earth have been dated at 3.8 billion years. Consequently, the geologic time scale runs from the present time to 3.8 billion years ago. Because the oldest rocks found in Idaho have been dated at 2.5 billion years, we have a

record of events in Idaho for a large part of the geologic time scale.

The present geologic time scale was originally developed by correlation of fossils based on relative age relationships. With the advent of radiometric dating techniques, the subdivisions of the time scale were assigned absolute ages. For example, we now know that the Triassic Period began about 245 million years ago.

Minerals

Minerals are naturally-occurring, inorganic (not derived from living things) compounds or elements. They have a definite chemical composition or range of compositions and a crystalline structure. Some substances are still considered minerals even though they are not covered by the definition given above. Among these are opal, a non-crystalline solid, and mercury, a liquid.

More than 2,500 minerals have been recognized. Their names are derived from languages such as Greek and Latin. Some are named for geographic localities in which they were first found and others are named after properties such as color, crystal form, density or change.

However, most commonly, they have been named after people.

Common Minerals

These 2,500 minerals do not occur in equal abundance; some are relatively common, but, most of them are rare. Only 10 elements occur abundantly in nature and they represent about 99 percent of the total mass of the Earth's surface. Thus the abundant minerals are composed of the abundant elements.

Subdivisions of Geologic Time and Symbols						
ERA	PERIOD AND SUBPERIOD		EPOCH	AGE (Ma)		
CENOZOIC	QUATERNARY (Q)		Holocene	0.01		
			Pleistocene	1.8		
	TERTIARY (T)	NEOGENE SUBPERIOD	Pliocene	5.3		
			Miocene	23.7		
			Oligocene	36.6		
		PALEOGENE SUBPERIOD	Eocene	57.8		
			Paleocene	66.4		
MESOZOIC	CRETACEOUS (K)		Late	144		
			Early			
	JURASSIC (J)			Late	208	
				Middle		
		TRIASSIC (T)			Early	245
					Late	
			Middle	286		
			Early			
PALEOZOIC	PERMIAN (P)		Late	286		
			Early			
	PENNSYLVANIAN (P)			Late	320	
				Middle		
		MISSISSIPPIAN (M)			Early	360
					Late	
	DEVONIAN (D)				Early	408
					Late	
		SILURIAN (S)			Middle	438
					Early	
	ORDOVICIAN(O)				Late	505
					Middle	
		CAMBRIAN (C)			Early	570
					Middle	
			Late	2500		
			Early			
PROTEROZOIC	NONE DEFINED			570		
ARCHEAN	NONE DEFINED			3800		

Geologic time scale. The age column is millions of years ago. Rocks older than 570 million years ago are also referred to as Precambrian.

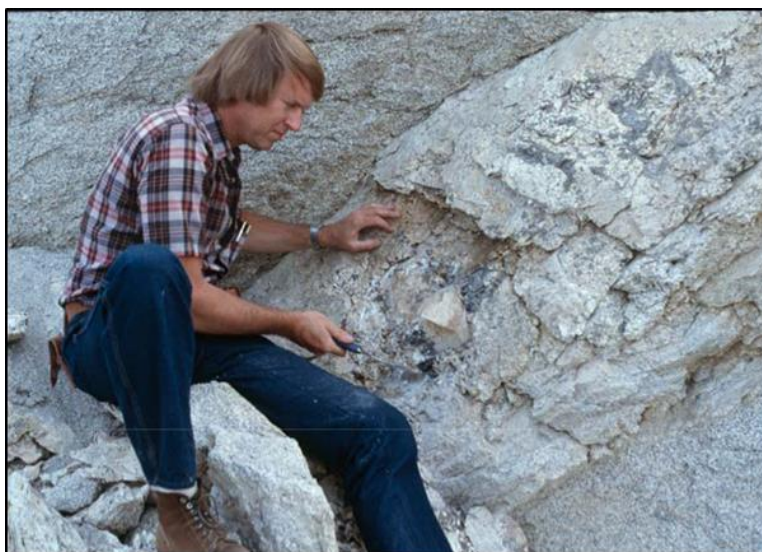
Rock-Forming Minerals

Rocks are aggregates of one or more minerals. For example, limestone is composed primarily of the mineral calcite. Granite typically contains three minerals: feldspar, quartz and mica. Certain minerals are so common in rocks they are called the rock-forming minerals. The minerals listed below make up most of the Earth and are so few in number that most beginning amateur prospectors or mineral collectors should be able to identify them. The important rock-forming minerals include quartz, potassium feldspar (orthoclase), plagioclase feldspar, muscovite mica, biotite mica, hornblende, augite, olivine, garnet, chlorite and clay. With the exception of olivine, all of the rock-forming minerals are common in Idaho.

Crystals

When crystalline minerals grow without interference, they have smooth, flat crystal faces that are related directly to the internal atomic structure of the mineral. Each mineral is assigned to one of six crystal systems. These systems are based on the number, position and relative lengths of the crystal axes.

Crystal axes are imaginary lines extending through the center of the crystal. The six crystal systems include isometric or cubic, tetragonal, hexagonal, orthorhombic, monoclinic and triclinic.



Geologist inspecting large feldspar crystal of pegmatite dike. The pegmatite dike is about 4 feet thick and was intruded into to granitic rock. Silent City of Rocks, Idaho.

Any given mineral crystal will grow in such a manner so as to form certain typical shapes or crystal habits. Crystal habits are used to identify minerals because they indicate the forms or combination of forms a mineral is likely to have. Cubic, columnar, and tabular are examples of crystal habits.

Mineral Identification

After a little practice the common rock-forming minerals can be identified on sight. However, some may require an examination of the various chemical and physical properties. The properties most useful in mineral identification include hardness, streak, color, specific gravity, fracture, cleavage, luster and shape or form. Some of the more important rock-forming minerals are discussed below.

The feldspar group constitutes the most important group of the rock-forming minerals. They are so abundant that they make up 60 percent of the Earth's crust. Feldspars are common in igneous, metamorphic and sedimentary rock.

- *Orthoclase* is the common potash (potassium) feldspar. It is transparent to translucent with colors of white, gray, yellow, pink, or colorless.
- *Plagioclase* feldspars are common in igneous rocks and some metamorphic rocks. Colors are white, yellow and gray. In much of the Idaho Batholith, plagioclase tends to be chalky white in color.
- *Quartz* is the second most common mineral and is widely distributed. Pure quartz is composed of silicon dioxide. It forms six-sided crystals with pyramidal ends. Colors include white, colorless, rose, purple, yellow, and smoky gray. Among the common crystalline varieties of quartz are amethyst (purple), milky quartz, rose quartz and smoky quartz. Very fine-grained varieties include agate, chalcedony, chert, flint and jasper. Some sedimentary rocks such as sandstone and quartzite are composed almost entirely of quartz.
- *Mica* has a perfect basal cleavage (also called micaceous cleavage) . Micas are one of the easiest minerals to identify because they consist of stacks of sheets or books of easily-parted plates.

Muscovite and biotite are the most common micas. Muscovite is the white, transparent mica and is most common in granites, and pegmatites. Biotite is the black mica and is common in both igneous and metamorphic rocks.

- The mineral *calcite* is composed of calcium carbonate. Calcite occurs primarily in sedimentary rocks and some metamorphic rocks such as marble. Calcite is the most abundant constituent of limestone. Many of the limestone formations in eastern Idaho consist of more than 90 percent calcite. Calcite has the property of reacting with cold dilute hydrochloric acid by effervescing or fizzing. Dolomite is calcium magnesium carbonate and reacts only mildly with acid.

Rocks

Rocks are naturally-formed, consolidated material composed of grains of one or more minerals. Geologists categorize rocks into three categories depending on their origin: igneous, sedimentary and metamorphic.

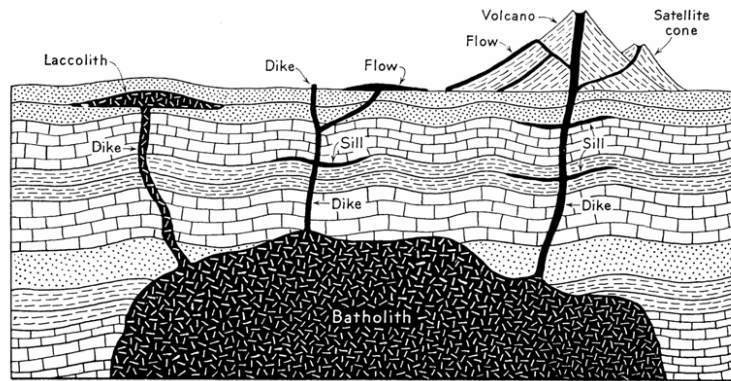
Igneous rocks are formed from solidification of molten material. Sedimentary rocks are formed by the accumulation of fragmental material derived from preexisting rocks of any origin as well as the accumulation of organic material or precipitated material. Metamorphic rocks occur as a result of high pressure, high temperature and the chemical activity of fluids changing the texture and (or) mineralogy of preexisting rocks.

Igneous Rocks

Magma

Igneous rocks are those rocks that have solidified from an original molten state. Temperatures within the Earth are so hot that many rocks and minerals are able to exist in a molten condition called magma. This molten rock exists deep below the Earth's surface in large pools called magma chambers. Many magmas or portions of magmas are lighter than the surrounding rock and tend to rise toward

the surface of the crust; also, the high pressure at depth facilitates the upward movement of magma. Molten materials that extrude through the surface of the Earth are called eruptive, extrusive or volcanic rocks. Those magmas that crystallize and solidify at depth, never reaching the Earth's surface before consolidation, are called intrusives or plutonic rocks. Of course after consolidation, plutonic rocks may be exposed at the Earth's surface by the process of erosion. The Idaho Batholith was emplaced as a number of overlapping plutons. These plutons consolidated into rock which was subsequently exposed by erosional processes.



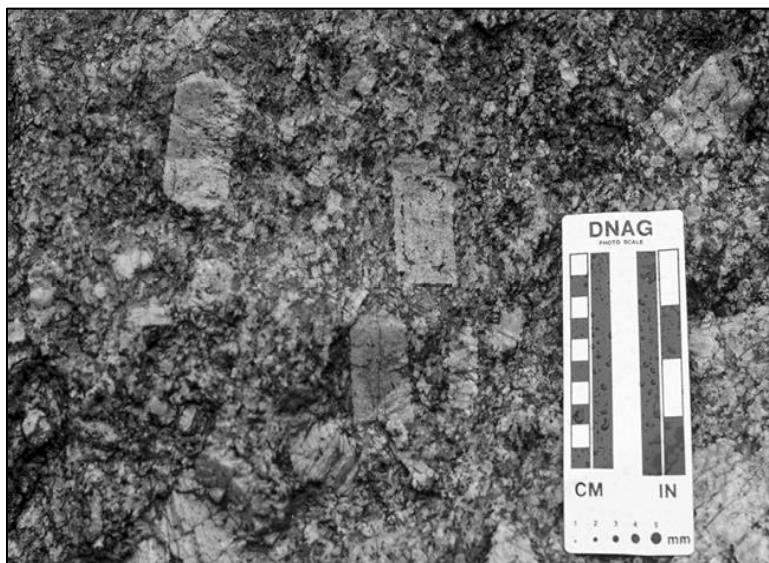
Cross section showing various types of plutons commonly found in the vicinity of a batholith.

The crystal size of igneous rocks is very diagnostic of their origin. Volcanic or extrusive rocks have a very small average grain size which is generally too small to discern with the naked eye. Extrusive rock has a very high component of glass because it was quickly frozen from the molten stage before crystals had time to grow. The more deeply-buried plutons cool more slowly and develop a coarse texture composed of large crystals. Therefore, large mineral crystals of more than one inch in diameter indicate formation at a depth of 6 to 12 miles.

Mafic and Felsic Magmas

Magmas are thought to be generated in the outer 60 to 180 miles of the Earth where temperatures are hot enough to cause melting. Magmas rich in magnesium, iron and calcium are called *mafic*. Those rich in sodium, potassium and silicon are called *felsic*. Those that are transitional between mafic and felsic are called *intermediate*. Felsic magmas are generated mostly within the continental crustal regions where the source of parent rocks are abundant;

whereas, mafic magmas may be derived from parent materials rich in magnesium, iron and calcium which occur beneath the crust. Mafic magmas, coming from a deep hot source, are about 1,200 degrees centigrade when they reach the Earth's surface; felsic magmas are much cooler—about 700 degrees centigrade upon reaching the Earth's surface.



Large rectangular orthoclase feldspar phenocrysts in a groundmass of smaller crystals. South of Stanley, Idaho.

Origin of Basalt

Most basalt originates at spreading centers such as the mid-oceanic ridge system and stationary mantle plumes. Basalt magma originates from partial melting of mantle material. The fluid magma rises through fissures formed by tensional forces of two diverging plates.

Origin of Andesite and Granite

Intermediate and felsic magma in Idaho are believed to have originated where a cool slab of oceanic lithosphere of basalt and overlying sedimentary rock descended beneath the continental crust of the western United States. The descending plate of lithosphere becomes hotter with increasing depth. Water trapped in the descending plate also lowers the melting temperature so that partial melting of basalt takes place. While the basaltic magma rises through the overriding continental crust, the magma absorbs some of the more silica rich rocks to become intermediate in composition. Also, the very hot basaltic magma chambers in the continental crust could melt the surrounding felsic rocks and create granitic magmas.

Emplacement of Magma

Bodies of intrusive rocks exist in almost every shape and size. Regardless of shape or size, they all come under the general term pluton. Most of them appear to be emplaced in the surrounding country rocks (host rocks) by the process of forceful injection. By forceful injection, the body is

intruded along zones of weakness, such as fractures, by pushing apart the surrounding rock. A pluton is also emplaced by melting rock around it and prying out blocks of the country rock. The surface between the pluton and the country rock is the intrusive contact. Magma is also aided in its upward movement because it is generally less dense than the surrounding rock. When the magma stops moving it begins to crystallize. Those plutons that reach shallow to intermediate depths tend to be porphyritic, that is, large crystals are contained in a finer crystalline groundmass.

Types of Plutons

Dikes are small tabular plutons which cut across layering in the host rocks. Dikes may range from one inch to tens of feet thick. They are much longer than wide and can commonly be traced a mile or more. Dikes are generally intruded along fractures and tend to have the composition of pegmatite, aplite (white, sugar-textured dikes) and basalt. In almost every roadcut through the Idaho Batholith of central Idaho, aplitic and pegmatitic dikes can be seen. Sills are also tabular bodies of the same approximate size and shape range as dikes. However, sills are concordant or parallel to the layers of the surrounding host or country rock.

The largest plutons consist of granite and diorite and are found in the cores of mountain ranges. The Idaho Batholith is a good example. A batholith is defined as a pluton with a surface exposure in excess of 40 square miles. If the exposure is less than that, the pluton is called a stock. It is commonly believed that buried batholiths underlie large areas of widespread silicic volcanics in Idaho. Many of the large batholiths such as the Idaho Batholith are known to be a composite of many granitic plutons.

Pegmatites

Pegmatite bodies have a relatively larger grain size than the surrounding igneous rocks. Individual crystals are known to reach more than 30 feet in length. A pegmatite may have the composition of a granite, diorite or gabbro. All three types are exposed in the large granitic plutons of Idaho. However, granitic pegmatites are by far the most common. In practically every exposure of granitic rock in the state, there are one or more granitic pegmatite dikes exposed. Although most of these pegmatites do not



Outcrop of quartzite has many sills parallel to the layering of the rock. The large dike cuts across both the layering of the rock and the sills; therefore, the dike is younger than the sills.

exceed 10 feet in thickness; an uncommonly large pegmatite more than 300 feet along its smallest dimension is exposed in the City of Rocks near the town of Oakley. The extremely large crystal size (generally 2 to 8 inches), is attributed to both slow cooling and low liquid viscosity. Pegmatites represent the last portion of a pluton to crystallize. These residual fluids are much richer in certain elements than the original magma. High amounts of silica and ions of elements that are necessary to crystallize sodium plagioclase and potassium feldspar must be abundant in the fluids. The fluids are also rich in certain elements that could not be used in the crystal structure of the previously crystallized minerals. Water is also very abundant which promotes slower cooling and a lower temperature of crystallization. Many pegmatites were intruded along existing fractures.

Most Idaho pegmatites are composed of orthoclase feldspar, quartz and muscovite. Careful inspection will also reveal small red garnets, black tourmaline and bluish-green aquamarine. Aquamarine is generally only found in the tertiary plutons.

Common Igneous Rocks

Igneous rocks are classified on the basis of their texture and composition. Although more than several hundred names have been given to igneous rocks, only a few major rock types are discussed below:

- *Granite* is the most common coarse-textured rock. It is formed at great depths within the Earth and has crystals ranging from microscopic to more than one inch in size. Granite typically contains quartz, feldspar, mica and hornblende. Granites are

generally light in color and may have a salt and pepper appearance. The feldspar may cause it to be white, gray, pink or yellowish brown. Most of the large bodies of plutonic rocks in Idaho have typical granitic texture and composition. Potassium feldspar and plagioclase feldspar make up most of the rock, though quartz may represent up to 25 percent of the bulk composition. The black minerals are commonly hornblende and biotite mica. Muscovite is also common in some granite.

- *Gabbro* is a dark, coarse-grained igneous rock. It is generally composed of plagioclase feldspar and augite. Gabbro is generally dark green or dark gray in color. Idaho has relatively little gabbro compared to granite.

Classification of Igneous Rocks

Texture	Composition		
	Quartz, Feldspar Hornblende, Biotite <u>Light Colored</u>	<u>Intermediate</u>	Feldspar, Augite <u>Dark Colored</u>
Coarse Grained	Granite	Diorite	Gabbro
Fine Grained	Rhyolite	Andesite	Basalt
Glassy	Pumice Obsidian		
Pyroclastic or Fragmental	Air – Fall Tuffs Ash – Flow Tuffs		

- *Pumice* is lava that solidified while gases were released from it. It is essentially a frozen volcanic froth. Because of the abundance of gas cavities, pumice is so light in weight that it can float in water. Pumice is generally light gray or tan and has the same chemical composition as obsidian, rhyolite and granite.
- *Diorite* is a coarse- to fine-grained plutonic rock and has a mineral composition that places it midway between granite and gabbro. It has little quartz or potassium feldspar. Diorite tends to be a gray rock due to the high amounts of plagioclase feldspar and iron-rich minerals.

Andesite is much finer grained than diorite but has the same mineral composition. Andesites are more common than rhyolites, but less common than basalts.

- *Rhyolite* is a volcanic rock with the same composition as granite. The major difference is its fine-grain size or glassy texture. Rhyolite is generally light colored and may be gray, white, tan or various shades of red. It has a characteristic streaked texture called flow banding. Flow banding is caused by slow flowage of highly viscous lava.
- *Obsidian* forms when magma of a rhyolitic composition cools so fast that crystallization of the minerals is not possible. Thus volcanic glass is essentially a frozen liquid. It is a lustrous, glassy black or reddish-black rock. Obsidian has a conchoidal fracture giving it very sharp edges.

Because of this property, it was commonly used to make tools and weapons by early Americans.

- *Basalt* is the fine-grained compositional equivalent of gabbro. It is by far the most abundant volcanic rock. For example, the volume of basalt in the Columbia Plateau is estimated to be 74,000 cubic miles. Basalt is normally coal black to dark gray when not weathered. Common constituent minerals include pyroxene, calcic plagioclase and olivine. Basalt commonly has small cavities called vesicles. Basalt flows are characterized by columnar jointing which causes polygonal vertical columns that look like giant fence posts stacked on end. Most of the large basalt flows are extruded from large fissures in the Earth's crust. Basalts are very common throughout Idaho, especially western and southern Idaho.

Porphyritic Texture

Some fine-grained rocks such as basalt, rhyolite, and andesite have a mixed texture of large and small grains. This texture is called porphyritic and is characterized by large crystals called phenocrysts surrounded by a groundmass (background) of smaller crystals.

Pyroclastic Rocks

In addition to the fluid lava extruded from a volcano, a great amount of lava is blown out the vent by violent gas explosions. All material driven out explosively is called pyroclastic. Large fragments such as spindle-shaped volcanic bombs fall near the vent. However, the dust-size fragments called ash are carried hundreds of miles by prevailing winds. Volcanic ash is composed of fragments of volcanic glass and small crystals. When air-fall ash deposits consolidate, they are called ash-fall tuffs. Excellent examples of most of these volcanic products can be observed at Craters of the Moon National Monument.

One type of pyroclastic rock very common in southern and east-central Idaho is the welded ash-flow tuff deposit. This material consists of a very hot mixture of fragments of pumice, cinders, crystals and glass shards, many of which are more than one inch in size. They blow out of the vent, collapse, and move rapidly down slope somewhat like a lava flow, but riding on a cushion of hot gases. When the deposit settles and comes together, the angular fragments are so hot they weld together. Unlike rhyolite flows, a single ash flow tuff unit may extend up to 100 miles. These tuffs make distinctive rim formers above the lake-bed deposits in the Snake River Plain.

Volcanic Cones

Volcanoes are vents in the Earth's crust through which molten rock and other volcanic products are extruded. There are three types of volcanic cones: cinder cones, shield volcanoes (lava domes), and composite cones (stratovolcanoes). All three types are common in southern Idaho.

- *Cinder cones* are formed entirely of pyroclastic material, mostly of cinders. These cones consist of a succession of steeply-inclined layers of reddened scoriaceous cinders around a central crater. They are generally less than 1,000 feet in height and are susceptible to erosion because there is generally nothing holding the mass together. This type of cone has the steepest flanks of the three types of volcanic cones. Hundreds of cinder cones are distributed throughout the Snake River Plain, generally aligned along fractures in the crust. These cones add much relief to the otherwise featureless plain.
- *Shield volcanoes* are built almost entirely of basaltic lava flows. They have gently-rounded profiles with a circular outline. This type of cone is the most stable and least susceptible to erosion.
- *Composite or stratovolcanoes* are composed of alternating sheets of lava and pyroclastic material. The alternating pyroclastic layers and lava layers indicate that the pyroclastic material was produced during periods of explosive activity, whereas the lava eruptions occurred at times of quiescence.
- *Calderas* are nearly circular basin-shaped depressions in the upper part of volcanoes.

They are much larger than volcanic craters and are generally more than 6 miles in diameter. Most of those in Idaho are thought to have formed by collapse caused by the sudden withdrawal of supporting lava. Such calderas are common in southern and east-central Idaho.

Sedimentary Rocks

Sedimentary rocks are derived from preexisting igneous, sedimentary and metamorphic rocks. These rocks contain many clues as to their origin and the conditions that existed while they formed. Sedimentary rocks make up 75 percent of the rocks at the Earth's surface but only 5 percent of the outer 10 miles of the Earth. Sediment, as distinguished from sedimentary rock, is a collective name for loose, solid particles and is generally derived from weathering and erosion of preexisting rock. After formation, sediments are transported by rivers, ocean waves, glaciers, wind or landslides to a basin and deposited. **Lithification** is the process of converting loose sediment into sedimentary rock and includes the processes of cementation, compaction and crystallization.

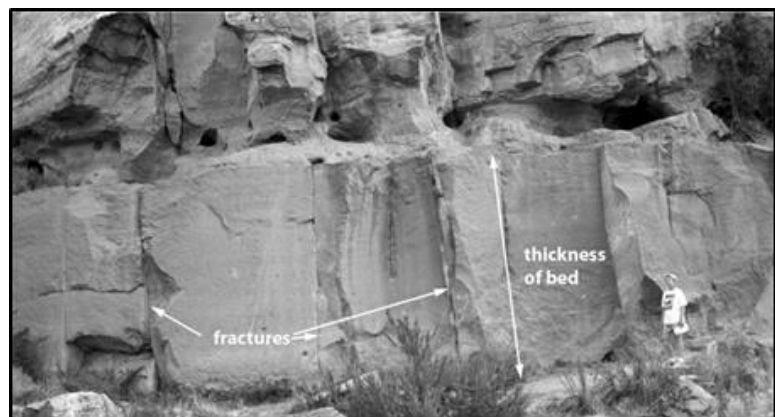
Sedimentary rock is formed by lithification of sediments, precipitation from solution and consolidation of the remains of plants or animals. Coal is an example of sedimentary rock formed from the compression of plant remains.

Rounding of Rock Particles

Rounding occurs during the transportation process by one or more of the erosional agents. Current and wave action in water are particularly effective in causing particles to hit and scrape against one another or a rock surface. The larger the particle the less distance it needs to travel to become rounded. For example, the boulders of the melon gravel deposited by the Bonneville flood were rounded after 3 to 6 miles of transportation. By contrast, a grain of sand may require hundreds of miles of transportation to become rounded.

Deposition of Sediment

Sorting of sediment by size is also effectively accomplished by moving water. A river sorts sediment by first depositing cobbles, then pebbles, sand, silt and clay. The larger the size of sediment, the greater the river's energy necessary to transport it. Deposition is the term used to describe the settling of transported



Sandstone deposited during the Pliocene in ancestral Lake Idaho. Table Rock, northeast of Boise, Idaho

sediment.

Lithification of Clastic Rock

Clastic or detrital sedimentary rock is composed of fragments of preexisting rock. The grains are generally rounded and sorted during the transportation process. Clastic sediment is generally lithified by cementation. Cementation occurs when material is chemically precipitated in the open spaces of the sediment so as to bind the grains together into a hard rock. Common cements include calcite, silica and iron oxides. A matrix of finer-grained sediments may also partly fill the pore space.

Common Types of Sedimentary Rock

- *Conglomerate* is the coarsest-grained sedimentary rock formed by the cementation of gravel-sized sediments. The gravel is generally rounded; however, it probably did not travel very far. Conglomerates are generally deposited by a river.

Origin		Particle Size/Composition	Rock Name
Detrital or Clastic	Boulder, Cobble, Pebble (> 2mm)		Conglomerate
	Sand (1/16 – 2mm)		Sandstone
		Silt and Clay (< 1/16mm)	Shale
Chemical	Inorganic	Calcite Dolomite Halite	Limestone Dolomite Salt
	Biochemical	Calcite Plant Remains	Limestone Coal

Simplified classification of sedimentary rock.

- *Sandstone* is a medium-grained sedimentary rock formed by the cementation of sand-sized sediments, with silt and clay forming the matrix. The material may be deposited by rivers, wind, waves or ocean currents.
- *Shale* is a fine-grained sedimentary rock composed of clay- and silt-sized fragments. Shale is noted for its thin laminations parallel to the bedding. Compaction is very important in the lithification of shales. Before compaction, the sediment may contain up to 80 percent water in the pore spaces.
- *Chemical Sedimentary Rocks* are formed by material precipitated from solution. Examples include rock salt, gypsum and limestone.
- *Organic Sedimentary Rocks* consist mostly of the remains of plants and animals. Coal is an organic rock formed from compressed plant remains. Idaho has a few coal deposits but none exists in commercial quantities.

- *Limestone* is a sedimentary rock composed of mostly calcite. Some limestones are chemical precipitates and other limestones consist mostly of grains of calcite or shells of marine invertebrates. The calcite grains in limestone recrystallize readily so as to form new and larger crystals.

Sedimentary Structures

Sedimentary structures in sedimentary rock are formed either during the deposition process or shortly after deposition. One of the most important structures is bedding. An important principle of geology holds that sedimentary rocks are deposited in horizontal layers. The bedding plane is the nearly flat surface separating two beds of rock. Bedding planes originate by a change in grain size, a change in grain composition or a pause in deposition during the depositional process.

- *Cross bedding* is characteristic of sand dunes, river deltas and stream channel deposits. It is most typically observed where sedimentary layers or laminations accumulate at a steep angle to the horizontal. Cross bedding forms from the downstream migration of ripples, bedforms and sand waves when particles are deposited on the front of a growing delta or migrating point bar. Cross bedding is a good indication of wind or current direction. Cross bedding is very common in the Lake Idaho sedimentary deposits, formations of the Belt Supergroup, as well as most stream deposits exposed along Idaho's many canyons.
- *Mud cracks* are sedimentary structures that are abundant in many of the formations of the Belt Supergroup as well as in many Paleozoic marine sedimentary formations in Idaho. Mud cracks are polygonal cracks formed in clay- and silt-sized sediments. They are caused by the exposure of lake bottoms, river bottoms and tidal flats to the sun after submergence in water. The cracks are caused by the sun drying and shrinking the upper several inches of the exposed mud flat.
- *Ripple marks* are small ridges, generally less than one inch high and 2 to 8 inches wide. The ridges are developed by moving water and form perpendicular to the direction of water movement. If the profiles of the ripple marks are symmetrical, they are caused by waves in lakes; if the profiles are asymmetrical, they are caused by currents. The steep sides occur in the down-current direction.

Metamorphic Rocks

Metamorphic rocks are those that have transformed from preexisting rock into texturally or mineralogically-distinct new rocks by high temperature, high pressure or chemically-active fluids. One or more of these agents may induce the textural or mineralogical changes. For example, minute clay minerals may recrystallize into coarse mica. Heat is probably the most important single agent of metamorphism. Metamorphism occurs within a temperature range of

100 to 800 degrees centigrade. Heat weakens bonds and accelerates the rate of chemical reactions. Two common sources of heat include friction from movement and intrusion of plutons. Pressure changes are caused primarily by the weight of overlying rock. Where there are more than 30,000 feet of overlying rock, pressures of more than 40,000 psi will cause rocks to flow as a plastic. Pressure may also be caused by plate collision and the forceful intrusion of plutons.



Steeply dipping angular outcrop of quartzite in eastern Idaho.

Chemically-active fluids (hot water solutions) associated with magma may react with surrounding rocks to cause chemical change. Directed pressure is pressure applied unequally on the surface of a body and may be applied by compression or shearing. Directed pressure changes the texture of a metamorphic rock by forcing the elongate and platy minerals to become parallel to each other. Foliation is the parallel alignment of textural and structural features of a rock. Mica is the most common mineral to be aligned by directed pressure.

Types of Metamorphism

There are two types of metamorphism: contact metamorphism and regional metamorphism. Contact metamorphism is the name given when country rock is intruded by a pluton (body of magma). Changes to the surrounding rocks occur as a result of penetration by the magmatic fluids and heat from the intrusion. Contact metamorphism may greatly alter the texture of the rock by forming new and larger crystals. In contact metamorphism, directed pressure is not involved so the metamorphosed rocks are not foliated.

Regional Metamorphism

Most metamorphic rocks in Idaho are caused by regional metamorphism. This type of metamorphism is caused by high temperature and directed pressure. These rocks are typically formed in the cores of mountain ranges, but may be later exposed at the surface by erosion. Typical rock types include foliated (layered) rocks such as slates, phyllites, schists and gneisses.

Common Metamorphic Rocks

- *Marble* is a coarse-grained rock consisting of interlocking calcite crystals. Limestone recrystallizes during metamorphism into marble.

Rock Type	Metamorphic equivalent under increasing temperature and pressure →				
	Slate	Phyllite	Schist	Gneiss	
Shale					
Sandstone				Quartzite	
Limestone				Marble	
Basalt			Schist	Amphibolite	
Granite			Granite	Gneiss	

Simplified diagram showing types of metamorphic rocks.

- *Quartzite* forms by recrystallization of quartz-rich sandstone in response to heat and pressure. As the grains of quartz grow, the boundaries become tight and interlocking. All pore space is squeezed out; and when the rock is broken, it breaks across the grains rather than around the grains. Quartzite is the most durable construction mineral. Although both marble and quartzite may be white to light gray, they may be readily distinguished because marble fizzes on contact with dilute hydrochloric acid, whereas quartzite does not. Also, marble can be scratched with a knife, whereas quartzite cannot.
- *Slate* is a low-grade metamorphic equivalent of shale. It is a fine-grained rock that splits easily along flat, parallel planes. Shale, the parent rock, is composed of submicroscopic, platy clay minerals. These clay minerals are realigned by metamorphism so as to create a slaty cleavage. In slate, the individual minerals are too small to be visible with the naked eye.
- *Phyllite* is formed by further increase in temperature and pressure on a slate. The mica grains increase slightly in size but are still microscopic. The planes of parting have surfaces lined with fine-grained mica that give the rock a silky sheen.
- *Schist* is characterized by coarse-grained minerals with parallel alignment. These platy minerals,



Strongly foliated (layered) gneiss with alternating layers of feldspar and biotite mica. West of Shoup, Idaho.

generally micas, are visible to the naked eye. Schist is a high-grade, metamorphic rock and may consist entirely of coarse, platy minerals.

- *Gneiss* is a rock consisting of alternating bands of light and dark minerals. Generally the dark layers are composed of platy or elongate minerals such as biotite mica or amphibolite. The light layers typically consist of quartz and feldspar. Gneiss is formed under the highest temperatures and pressures which cause the minerals to segregate into layers. In fact, slightly higher temperatures than necessary to convert the rock into gneiss would cause the rock to partially melt.

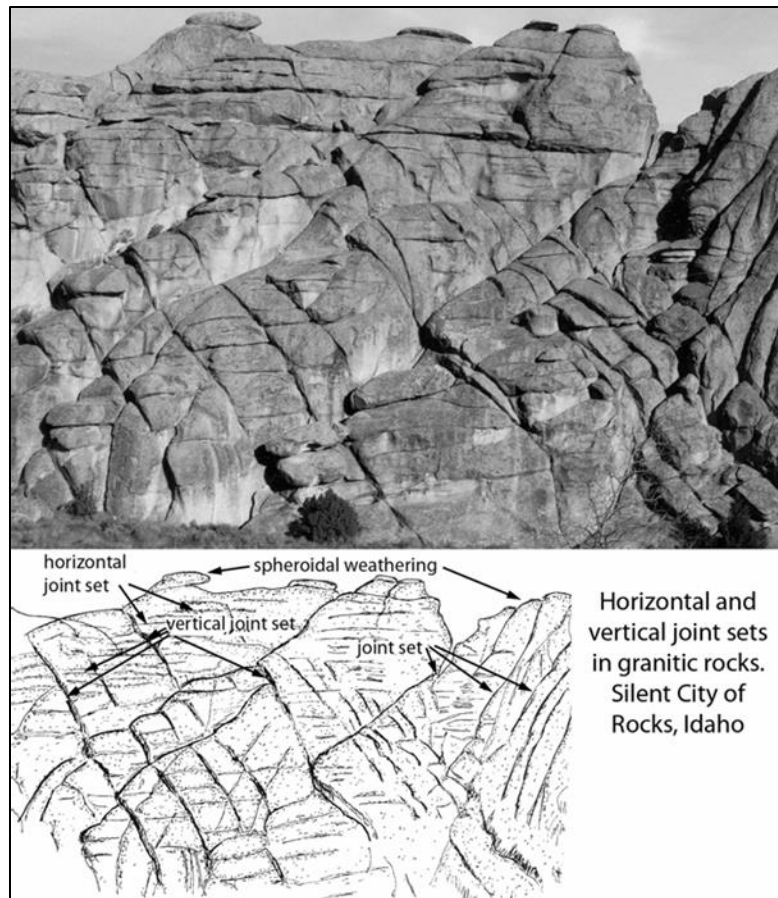
Structural Geology

The crust of the Earth is constantly moving. However, with the exception of faults accompanied by Earthquakes, this rate of movement is far too slow to notice. In the mountain ranges of Idaho, movement generally occurs at a much higher rate than it does in the more stable interior of the continent.

Folds

Folds are bends in rock layers generally caused by compression. Typically there are a series of arches (upfolds) and troughs (downfolds). This type of deformation is plastic so the rocks were probably buried deeply in the Earth's crust when the folding occurred. High temperatures and pressures deep in the crust allow rocks to deform as a plastic rather than break. On the other hand fractures such as faults and joints occur near the surface where the rock is cold and brittle. Therefore you can see that the type of deformation, plastic or fracture, indicates the level in the crust where the deformation occurred.

Several terms are necessary to describe and interpret a series of



folds. An anticline is an upfold or arch and where layers dip away from the axis (or hinge line). A syncline is a downfold or arch. Synclines and anticlines are typically plunging folds. In a plunging fold the axes are not horizontal. In a dome, the beds dip away from a central point and in a structural basin the beds dip towards a central point.

Fractures

If a rock is brittle, it may rupture or break under stress. Most rock near the Earth's surface is brittle so almost every exposure of bedrock is cut by fractures. There are two types of fractures in rock: joints and faults. A joint is a fracture along which no movement has taken place. Joints are generally caused by tensional forces. A fault is a fracture or break in the rock along which movement has taken place. The rupture and subsequent movement may be caused by tensional, compressional or shear forces.

Joints

Joints are fractures in rock where no displacement has occurred along the fracture surface. Columnar jointing is a specialized type of jointing common to volcanic flows. Hexagonal columns form in response to contraction of a cooling lava flow. Exfoliation (or sheeting) is another specialized type of joint generally caused by expansion parallel to the weathering surface. Where closely-spaced joints are parallel, they make up a joint set. These joints may be spaced from several inches to tens of feet apart. Typically rock exposures exhibit two or more joint sets. The study of joints is important for site evaluations for dams or other facilities because jointing can affect the permeability and strength of the rock. Joints are also important as a plumbing system for hot water convection and the emplacement of mineral deposits.

Faults

Faults are fractures in rock along which movement has taken place parallel to the fracture plane. Many faults are active, that is, movement has taken place during historical times. Where faults are exposed in bedrock the geologist looks for evidence of displacement or offset features to determine the amount of displacement and the relative direction of movement. Fault planes or zones vary considerably in thickness. Some are just a thin crack in the rock, whereas others may consist of a brecciated and



Looking at the upper block of a fault with large vertical grooves or striations on the upper fault surface. West of Ketchum, Idaho.

sheared zone up to 1,000 feet wide. Faults also range in length from several feet or less to hundreds of miles. For example the San Andreas fault extends about 620 miles through western California, slowly moving Los Angeles toward San Francisco. The current rate of movement averages about one inch per year so it will take about 25 million years to make Los Angeles a western suburb of San Francisco. During the 1906 Earthquake that devastated much of San Francisco, bedrock along the fault was displaced as much as 15 feet. The three major types of faults include normal or gravity faults, reverse or thrust faults, and strike-slip or transcurrent faults.

- A *normal fault* is one along which the hanging wall has moved down relative to the footwall. The fault plane of normal faults typically dips at an angle of 60 degrees from the horizontal. The normal fault is the most common type of fault that you can expect to see in the field. The largest and most impressive group of normal faults are those that form the fault blocks that make up the Basin and Range Province of eastern Idaho. Normal faults are caused by rupture in response to tensional forces. Because the rock is pulled apart rather than pushed together, the broken area has much space available for ore solutions to move in and precipitate. Most lode or vein deposits are formed in normal fault zones.
- In a *reverse fault*, the hanging wall moves up relative to the footwall. The fault plane is typically inclined 30 degrees from the horizontal, but may vary significantly from this. Reverse faults are not as common as gravity or normal faults.
- A *thrust fault* is a type of reverse fault that is characterized by a low angle of inclination of the fault plane. In fact the fault plane is commonly horizontal or subhorizontal. Both reverse and thrust faults are caused by rupture in response to compressional forces. Eastern Idaho has many exceptional examples of large thrust faults where the upper plate has moved from west to east tens of miles placing older rocks over younger rocks.
- A *strike-slip fault* is one along which the movement has been parallel to the strike of the fault plane and is caused by rupture in response to shear forces. If an observer looks along the strike of a left-lateral, strike-slip fault, the relative movement has been such that the left-hand side has moved towards the observer. Along a right-lateral, strike-slip fault, the block on the right has moved towards the observer.

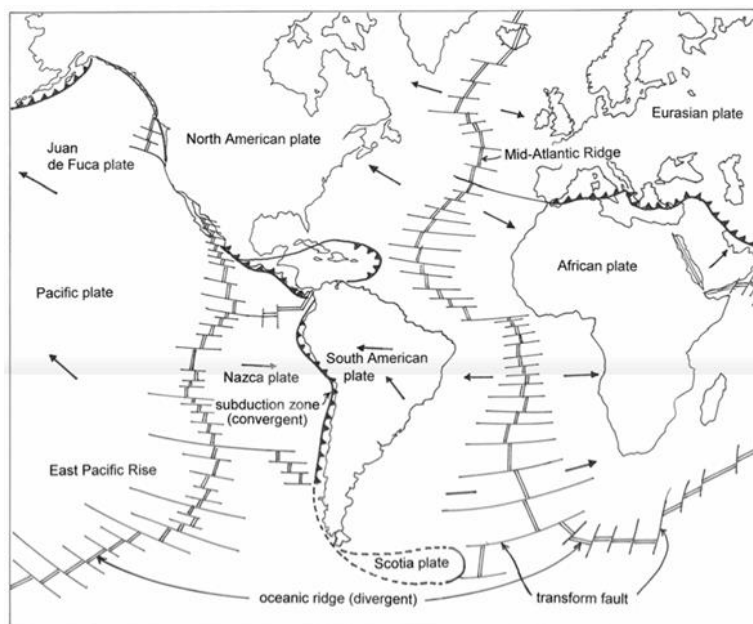
Plate Tectonics

The theory of plate tectonics has revolutionized the thinking of geologists. This is a unifying theory that explains many seemingly unrelated geologic processes. Plate tectonics was first seriously proposed as a theory in the early 1960s although the related idea of continental drift

was proposed much earlier.

The Plates

The outer part of the Earth is broken into rigid plates approximately 62 miles thick. These outer plates are called lithosphere and include rocks of the Earth's crust and upper mantle. Below the rigid lithosphere is the asthenosphere, a zone around the Earth that is approximately 90 miles thick and behaves like a plastic because of high temperature and pressure. The lithosphere plates move over the plastic asthenosphere at a rate of an inch or more per year. Eight large plates and a few dozen smaller plates make up the outer shell of the Earth.



The 8 major plates with their relative movement directions shown on the sketch.

The internal heat of the Earth is the most likely cause of plate movement; this heat is probably generated by the decay of radioactive minerals. The entire surface of the Earth is moving, and each plate is moving in a different direction than any other. We now believe that plate movement is responsible for the highest parts of the continents and the deepest trenches in the oceans. Such movements also cause catastrophic events like Earthquakes, volcanoes and tsunamis.

Plate Boundaries

Plate boundaries are of three types: a diverging plate boundary is a boundary between plates that are moving apart; a converging plate boundary is one where plates are moving towards each other; and a transform plate boundary is one at which two plates move past each other.

Diverging Boundaries

Diverging boundaries occur where plates are moving apart. Most of these boundaries coincide with the crests of the submarine mid-oceanic ridges. These ridges form by ascending hot mantle material pushing the lithosphere upward. When heat rises, molten rock moves upward and the expansion from the heat and pressure causes the ridge plate to bow upward and break apart at the spreading centers. Tension cracks form parallel to the ridge crest and molten rock from magma chambers in the mantle is intruded through the fractures. Magma erupts into submarine

volcanoes and some of it solidifies in the fissure. New crust forms in rifts at the spreading centers. As new magma is extruded, it accretes to both sides of the plates as they are pushed or pulled apart. As the plates continue to pull apart, new tension fractures form and fill with magma. This cycle repeats itself again and again.

Transform Boundaries

The transform boundary occurs where two plates slide past one another. The San Andreas Fault is one of the best known land exposures of a transform boundary.

Converging Boundary

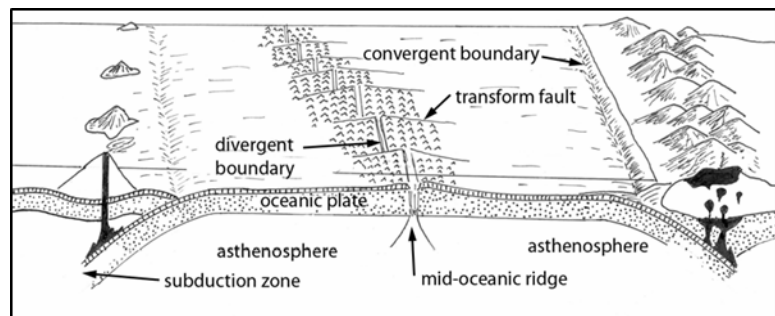
A converging boundary, where plates move toward each other, is responsible for the origin of most of Idaho's igneous rock as well as most of the major structural features of the state. Where one plate is covered by oceanic crust and the other by continental crust, the less dense continental plate will override the denser oceanic plate.

The older the oceanic plate, the colder and more dense it is. Where two plates collide, the dense plate is subducted below the younger and less dense plate margin. At this boundary, a subduction zone forms where the oceanic plate descends into the mantle beneath an overriding plate. As the oceanic plate descends deeper into the Earth it is heated progressively hotter. Also the friction caused by the two plates grinding past each other leads to greater temperatures.

At the subduction zones, submarine trenches form, representing the deepest parts of the ocean basins. Earthquakes continuously occur at the plate margins where the overriding plate is grinding and abrading the subducted layer. The subducted plate causes Earthquakes all along its downward path as it slowly moves into the Earth's mantle. By measuring the depth and position of the Earthquakes, geologists are able to determine almost exactly the position and orientation of the subducting plate.

Generation of Magma, Volcanoes and Batholiths

When the plate reaches a certain depth, heat and pressure melt the lighter minerals within it. This light molten rock or magma coalesces at depth and floats upward through the more dense rock towards the surface of the Earth. Where these globs of molten rock break through the oceanic crust, they form chains of volcanic islands. The Aleutian Islands are a well-known example.



Cross section of the asthenosphere and the lithosphere showing three types of plate boundaries: (1) a divergent-plate boundary at the mid-oceanic ridge; (2) an oceanic-oceanic convergent plate boundary on the left side; and (3) and oceanic-continental convergent plate boundary on the right side

The portion of the magma that manages to break through the surface and forms volcanoes is classified as volcanic or extrusive rock. The portion that does not break through the Earth's surface, but instead solidifies within the Earth's crust, is classified as intrusive igneous or plutonic rock.

Where an oceanic plate is subducted below a continent, it partially melts and the rising globs of magma melt and absorb portions of the silica-rich, low specific gravity continental rocks. Where magma manages to break through the continental crust, the extrusive products are much more siliceous than their oceanic counterparts.

Mantle Plumes

Mantle plumes are believed to form where convection currents in the Earth's mantle cause narrow columns of hot mantle rock to rise and spread radially outward. One of the most convincing theories for the origin of the Snake River Plain proposes that a hot mantle plume left a track across the plain from west to east and was the source for most of the volcanism in the Snake River Plain. The hot mantle plume is stationary and the overriding lithospheric plate moves slowly in a west-southwest direction. This hot mantle plume is now thought to underlie the caldera at Yellowstone Park. Many of the volcanic islands of the Pacific Ocean may have originated from a mantle plume. The best known examples are the Hawaiian Islands.

Origin of Mountains

Mountains are the most conspicuous landforms in Idaho. Any isolated mass of rock may be called a mountain because no minimum height or shape is required. Mountains may be formed by volcanoes, by erosional processes and by structural processes such as faulting and folding.

Volcanic Mountains

There are many volcanic mountains in southern Idaho, particularly within the Snake River Plain Province. These mountains generally consist of individual cones of cinder and extrusive igneous rock. The volcanic material was extruded through a central vent in the Earth's crust and piled up on the surface to form a cone. In Idaho, volcanic mountains tend to be smaller than other types of mountains, generally less than 1,000 feet high. Most of the volcanic mountains are more properly referred to as hills (a mountain is generally defined as projecting more than 1,000 feet above the surface; if the cone does not reach that height, it should be referred to as a hill). Volcanic hills or mountains also tend to be isolated and erratically distributed, although they are commonly aligned along rifts or fissures such as the Great Rift of the Snake River Plain. They are generally dome to conical shaped and are symmetrical in plain view. As a general rule, volcanic mountains consisting mostly of cinders and tuffaceous material are the most susceptible to erosion of all mountains.

Erosional Mountains

Erosional mountains are found in regions of crustal uplift such as the central Idaho uplands. They are characterized by steep gorges, precipitous slopes and youthful streams. Idaho's erosional mountains have primarily been carved by glaciers and running water and are the result of hundreds of thousands of years of erosion in the intervening valleys.

Structural Mountains

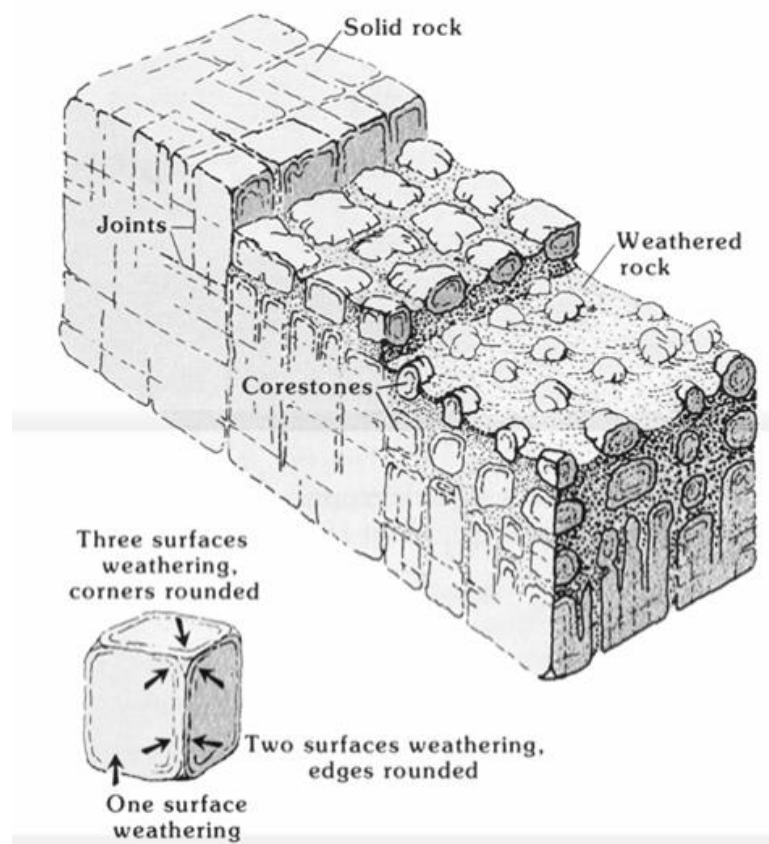
Structural mountains were created by structural activity such as folding and faulting. The Basin and Range Province of eastern Idaho is an outstanding example of mountains created by faulting. In the Basin and Range Province, large elongate blocks of the Earth's crust were moved up relative to the intermontane valleys along large normal faults. Once the ranges were moved up, erosional processes of primarily water and ice shaped the mountains into their present form.

Weathering and Soils

Weathering causes the disintegration of rock near the surface of the Earth. Plant and animal life, atmosphere, and water are the major causes of weathering. Weathering breaks down and loosens the surface minerals of rock so they can be transported away by agents of erosion such as water, wind, and ice. There are two types of weathering: mechanical and chemical.

Mechanical Weathering

Mechanical weathering is the disintegration of rock into smaller and smaller fragments. Frost action is an effective form of mechanical weathering. When water trickles down into fractures and pores of rock and then freezes, its volume increases by almost 10 percent. This causes outward pressure of about 30,000 pounds per square inch at -7.6 Fahrenheit. Frost



Weathering of joint blocks and stages in development of corestones. The corners and edges of the granite blocks are attacked more readily by weathering along joints, and rounded corestones result.

action causes rocks to be broken apart into angular fragments. Idaho's extreme temperature range in the high country causes frost action to be a very important form of weathering.

Exfoliation is a form of mechanical weathering in which curved plates of rock are stripped from rock below. This results in exfoliation domes or dome-like hills and rounded boulders.

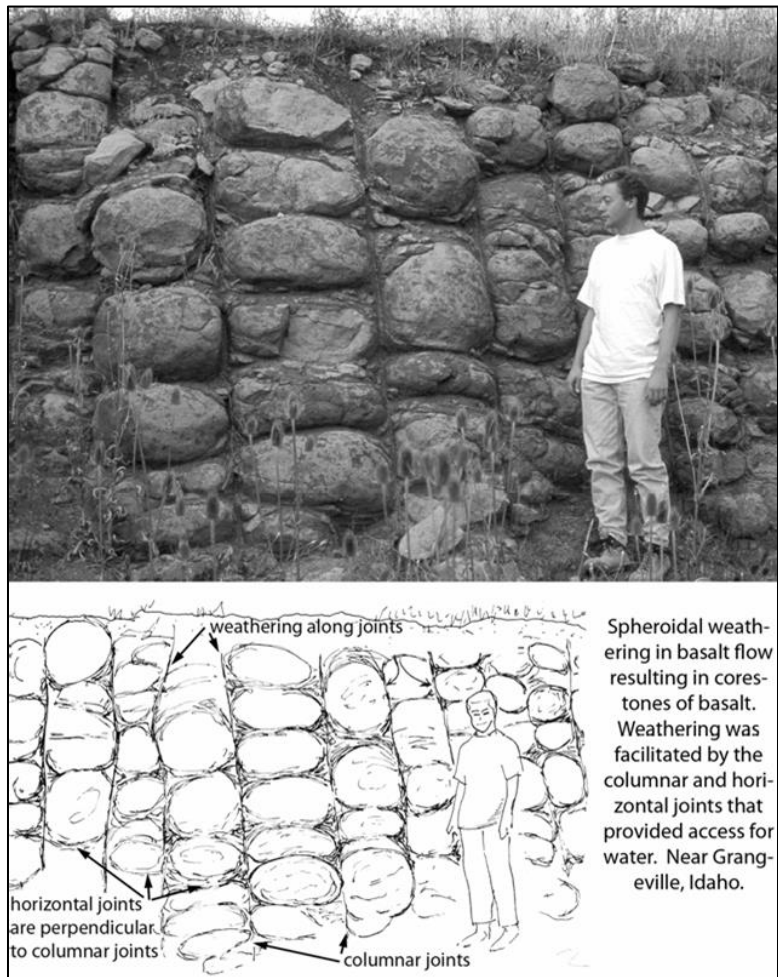
Exfoliation domes occur along planes of parting called joints, which are curved more or less parallel to the surface. These joints are several inches apart near the surface but increase in distance to several feet apart with depth. One after another these layers are spalled off resulting in rounded or dome-shaped rock forms. Most people believe exfoliation is caused by instability as a result of drastically reduced pressure at the Earth's surface allowing the rock to expand.

Exfoliation domes are best developed in granitic rock. Yosemite National Park has exceptional examples of exfoliation domes. Idaho has good examples in the Silent City of Rocks near Oakley as well as in many parts of the granitic Idaho Batholith. In fact, these characteristic rounded forms make granitic rock exposures easy to identify.

Another type of exfoliation occurs where boulders are spheroidally weathered. These boulders are rounded by concentric shells of rock spalling off, similar to the way shells may be removed from an onion. The outer shells are formed by chemical weathering of certain minerals to a product with a greater volume than the original material. For example, feldspar in granite is converted to clay which occupies a larger volume. Igneous rocks are very susceptible to mechanical weathering.

Chemical Weathering

Chemical weathering transforms the original material into a substance with a different composition and different physical characteristics. The new substance is typically much softer and more susceptible to agents of erosion



than the original material. The rate of chemical weathering is greatly accelerated by the presence of warm temperatures and moisture. Also, some minerals are more vulnerable to chemical weathering than others. For example, feldspar is far more reactive than quartz.

Differential weathering occurs when some parts of a rock weather at different rates than others. Excellent examples of differential weathering occur in the Idavada silicic volcanic rocks in the Snake River Plains. Balanced Rock and the Gooding City of Rocks are outstanding examples of differential weathering where hard layers alternate with soft layers

Features of Soil

Soil forms a porous mantle on the Earth's surface and acts as a reservoir for accumulating and storing water from snow and rain. When the water is released, it is done at a slow, steady dependable rate to charge streams and rivers. Soils are not only crucial for sustaining plant growth, but also are important for flood control.

The field study of soils is necessary in order to learn the potential and limitations of soil at a given site. A soil survey will give basic information about soil resources essential for land-use planning as it relates to development of undeveloped lands or conversion of land to new uses. This information will indicate the limitations and potential of a soil for recreation, planning, grading, erosion control, and suitable vegetation. For agriculture uses, soil information will aid in determining the appropriate irrigation system, including the length of run, water application rate, soil deficiencies, soil amendment requirements, leaching potential, drainage requirements, and proper field practices for maintaining optimum plant growth. Soil surveys also give information on sources of topsoil, and sand and gravel resources. Information on the engineering performance of soils is crucial to the construction industry. For example, engineers need to know how well a particular soil will support a building, the kinds of subgrades required for streets and roads, and whether onsite waste disposal systems will function properly.

Soil: a Definition. The term soil has different meanings to different people, depending on the context of its use. Most commonly, it is the natural medium for the growth of plants, regardless of whether it has discernible soil horizons. Soil is defined by the Department of Agriculture as a natural body comprised of solids (minerals and organic matter), liquid, and gases that occurs on the land surface, occupies space, and is characterized by one or both of the following: horizons, or layers, that are distinguishable from the initial materials as a result of additions, losses, transfers, and transformations of energy and matter or the ability to support rooted plants in a natural environment. Soil consists of the horizons near the Earth's surface that, in contrast to the underlying parent material, have been altered through time by climate, relief, and living organisms. Many properties of soil change with the seasons, including the pH, soluble salts, carbon-nitrogen ratio, biological activity, numbers of organisms, temperature, and moisture.

In the development of soil, the decomposed parent material (rock or alluvium) differentiates vertically with depth into layers or horizons. The soil profile is a vertical section from the surface of the soil through all horizons into the fresh parent material. The amount of time necessary to develop the profile depends on the weathering processes and the parent rock. These horizons are developed by the following processes:

1. Organic material accumulates at the surface.
2. Parent material is leached so that one or more minerals and/or weathering products are removed. Leaching is the removal of materials in solution by water percolating through the soil.
3. Organic material, including humus, accumulates in the upper portions of the soil profile. Humus is the dark-colored, fairly stable portion of the organic material that remains after the accumulated plant and animal residues decompose.
4. The weathering products accumulate in the lower part of the soil profile through the process of leaching and eluviation. *Eluviation* is the removal of soil material in suspension from upper layers (the zone of eluviation). *Illuviation* is the deposition of colloidal-size soil material moved from an upper horizon to a lower horizon in the soil profile (the zone of illuviation).

A soil horizon is a layer, approximately parallel to the surface of the soil, distinguishable from adjacent layers by a distinctive set of properties produced by soil-forming processes. The term layer, rather than horizon, is used if all the properties are believed to be inherited from the parent material or no judgment is made as to whether the layer is genetic.

The solum (plural, sola) of a soil consists of a set of horizons that are related through the same cycle of pedogenic processes. In terms of soil horizons, a solum consists of A, E, and B horizons and their transitional horizons and some O horizons. The solum of a soil presently at the surface includes all horizons now forming. Solum and soils are not synonymous. Some soils include layers that are not affected by soil formation; however, these layers are not part of the solum.

Residium is a term used when the properties of the soil indicate that it has been derived from the underlying rock and there is no evidence that it was transported into place. For example, if the rock fragment distribution decreases in quantity with depth, the material has likely been transported downslope as colluvial material. Also, if the rock fragments in the soil are of a different lithology than the underlying bedrock, they most likely would have been transported downslope. Movement of soil downslope is an important process in mass wasting; even where slopes are gentle, significant downslope soil movement can take place.

Soils consist of solids, liquids and gases. The solid component includes both small mineral and organic particles. It also has a large interconnected pore space shared by both water and air. The water and air ratios change constantly due to the loss of water by plants, evaporation and, recharge by rain and irrigation. Mineral particles, which are derived from the weathering of rock, are typically represented by 3 sizes:

Sand	0.05 - 2.0 mm
Silt	0.002 - 0.05 mm
Clay	< 0.002 mm

The relative amounts of sand, silt and clay determine the texture of the soil and its ability to transfer air and water. The organic component of soil results from the decomposition of plant material. Humus is a dark-colored material formed from the decomposition of plant and animal residues. It is composed of altered organic compounds formed from plants as well as additional compounds formed by organisms involved in the decay process. Clay and humus particles typically have a negative charge which allows them to absorb cations (positively charged ions) from solutions. Water takes nutrients in solution from minerals and organic material and transfers the nutrients to the plant roots.

Soil Formation Factors

There are five factors that influence soil development: parent material, climate, living matter, topography, and time. These factors operate both independently and interdependently.

- *Parent Material.* Parent material is the original geologic material that has been transformed into soil. Parent materials can be bedrock (residuum) such as granite, sandstone, limestone, or basalt. Other parent materials may be transported deposits carried to the site of deposition by water, ice or wind. Examples of transported materials include loess, dune sand, glacial till, volcanic ash, and lake deposits. The texture and mineralogy of the parent material have a large bearing on the relationships to the other factors. Texture has a large role in determining porosity and permeability for water flow and storage which in turn influences plant growth, organic material accumulation, weathering, and leaching. The mineralogy determines the weathering rate and weathering products in a given environment. For example, carbonate-bearing material weathers rapidly from leaching, while siliceous material such as quartz weathers very slowly. Soils may closely resemble parent material if there has been a lack of time such as may be the case with glacial deposits, sand dunes, and other alluvial deposits.
- *Climate and Soil Formation.* Climate affects soil most directly through temperature and precipitation. The amount of water entering the soil plays a large role in the growth of plants. Also, water facilitates a great many chemical and physical processes

that enhance weathering. With sufficient water, the rate of mineral weathering increases with temperature. The tropical regions which are both warm and wet are subject to intense weathering. However, high temperature causes rapid decay and disappearance of organic residues from the soil. High precipitation or irrigation also causes leaching, the removal of soil materials, including chemicals and nutrients by water flowing through the soil.

- *Living Material.* Plants supply a protective cover that retards erosion and enhances infiltration of water into the soil. Plants are the primary sources of organic material in the soil. Plant roots help break apart rocks, mix soil materials, form soil structure, absorb nutrients and ultimately provide residues. Root channels also provide access for water and air movement through the soil. Many organism, including rodents, earthworms, and insects assist decomposition through burrowing and mixing. Mixing by animals transports raw plant debris from the soil surface down into the topsoil. Microscopic organisms are important for organic material decay and humus formation. They also release nitrogen from proteins in the organic material. Nitrogen is an essential nutrient that plants need in large quantities. Microorganisms and the humus they produce act as a glue to hold soil particles together in clumps or aggregates. Well-aggregated soil facilitates the movement of air and water to plant roots.
- *Topography.* Topography causes localized changes in such factors as parent material, moisture, and temperature. Topography influences soil formation by controlling runoff and the effectiveness of precipitation. Topography also determines to a significant degree the climate. The direction and the steepness of the slope also determine how effectively the sun can warm the soil and cause evaporation. Soils on north-facing slopes tend to be cooler and wetter than soils on south-facing slopes. The drier and warmer south-facing slopes generally produce less vegetation and therefore have lower organic matter content and are more subject to erosion. Areas of low relief can retain much more precipitation than can areas on steep slopes. Steep slopes are also susceptible to mass movement and creep which limit soil development.
- *Time.* Many soils are thousands and even tens of thousands of years old. The soil-forming factors discussed above contribute the total environment for soil formation. They establish the types of changes and the rate of change. There is a great difference in the rate and kinds of reactions, depending on the environment of formation. Of all the soil-forming processes, weathering occurs at the slowest rate. For example, decomposition of siliceous material may take tens of thousands if not hundreds of thousands of years. Of course, in warm moist climates, weathering occurs at a much more rapid rate. Soils tend to have the same color, texture, and chemical composition as their parent materials. The older the soil, the less impact parent material has on soil

characteristics. As soils age, many original minerals are destroyed and many new ones are formed. Soils tend to become more leached, more clayey, and more acid with increasing time.

Soil Formation Processes

Properties contributed by the parent material are called inherited; acquired properties are created by chemical and physical changes to the parent material. Acquired properties, include a new texture, color, or structure. Soil-forming processes include a combination of physical, chemical, and biological reactions:

- *Mineral weathering and clay development.* Coarse grains are broken down to clay-size particles. This process is most significant in warm, wet climates such as in the tropics. Clay soils may be very deep. Also, most climates enhance leaching or removal of soluble materials by water migrating downward. This causes tropical soils to be infertile.
- *Accumulation of organic material.* Organic material is formed by the decay of plant residue. Trees contribute residues with leaves, needles and twigs. Microorganisms that live in the soil feed on fresh organic matter and change it into humus.
- *Ion Exchange.* Ions from the soil solution displace ions absorbed to clay and humus. This process is less important in arid regions.
- *Translocation* occurs where a component of the soil is moved to a different level by water. In most cases this process involves the eluviation or movement of a component near surface to redeposition or illuviation, in the horizon below it. Water moving through the soil transports very fine clay particles from one horizon to another, or from place to place within a horizon. When the water stops moving, the clay particles are deposited on the surface of the soil aggregates. These coatings of clay particles are called clay skins, which have a distinctive dark, waxy appearance. Lime and gypsum are commonly translocated this way in arid regions. In low rainfall areas, leaching is typically incomplete. Water moves down through the soil profile, dissolving soluble minerals on the way. But because there is insufficient water to move the minerals completely through the soil profile, when the water stops moving due to surface evaporation or plant use, the salts are precipitated. The most common deposits of this type are the well know caliche layers found throughout southern Idaho. When these layers become cemented with lime and/or silica, they form hardpans or duripans. Upward translocation may occur in places where there is a high water table for extended periods of time. Evaporation at the surface causes water to move upward continually so that salts are dissolved on the way up through the profile and are precipitated as the water evaporates.

- *Structure formation* occurs where physical changes separate a cohesive soil into an aggregate of various shapes and sizes. Expanding plant roots or ice crystals may break up the cohesive layer, or shrinkage of the mass may leave dessication cracks.
- *Mixing of soil* is most commonly accomplished by burrowing insects and worms.
- *Precipitation* may add about 5 pounds of nitrogen every year to each acre of soil. When soils are very wet, nitrogen can be changed to a gas and lost to the atmosphere. Precipitation can be acid and facilitate leaching and chemical changes. Precipitation also causes soils to erode and rivers to flood.

Features that Affect Soil Quality

- *Organisms in Soil.* Many organisms including plants and animals and microscopic organisms live within soil. Many are beneficial such as combining carbon and other nutrients into organic compounds, and others facilitate decomposition of plant and animal residues. Also, the physical condition of soil is enhanced by the burrowing and mixing by rodents, insects, and worms. However, some organisms, such as parasites, are not beneficial. Viruses are also undesirable because they cause diseases. The causes of diseases in plants include bacteria, fungi, actinomycetes and viruses. Decay and erosion both diminish the organic material in soils. This reduction must be offset by continuous replenishment of fresh residues. Unfortunately, in most soils, the loss of organic material through erosion and decay is not balanced by the addition of organic material in residues.
- *Aeration of Soil.* The aeration of soil is the process by which air in the soil is replaced by air from the atmosphere. The most common cause of poor aeration is waterlogged soils with no air access.
- *Water.* Water is crucial for all biological, chemical and physical processes in soil. Because it is also essential for plant growth, a very important function of soils is to store and convey water to the plant roots. The rate of water flow through soil is determined by the gradient of the potential and the hydraulic conductivity of the soil. The conductivity of water through soil depends on the texture and the saturation of the soil. The rate of water flow through soils is highest if the soils are coarse textured and the pores are filled with water. However, if the soils are unsaturated, fine-textured soils are the best conductors of water.
- *Available water* is the water stored in the soils for use by plants. Available water, which depends on soil depth and texture, is estimated by the difference in the amount

of water a soil holds at field capacity (the most water a soil can store after wetted and drained until drainage ceases) and the permanent wilting point. Coarse-textured soils hold the least amount of water in available form, medium-textured soils store the most amount of water, and fine-textured soils store an intermediate amount of water.

- *Soil Erosion.* Soil erosion is caused by wind or water. Examples are agricultural practices that leave soil exposed are overgrazing, tilling, and logging. Factors that determine the extent of rainfall erosion include climate, soil, slope, nature of the crop, and any erosion control measure that is used. The surface soil, which is generally the soil material lost, is the most productive portion of the soil. The less productive soils are typically more susceptible to erosion; so, the more a soil is eroded, the more susceptible it becomes to erosion. Wind erosion factors include soil, soil surface roughness, climate, the type and extent of plant cover, and the field width.
- *Nutrient Deficiencies.* Nutrient deficiencies in plants are identified by symptoms such as off- colored or poorly developed leaves (may be chlorosis), stunting, deformed fruit and cracking of stems. Nitrogen, phosphorous, and potassium are required by plants in large amounts. Plants rely on soil to furnish the minerals; however, soils are typically deficient in certain minerals. Nitrogen and phosphorous are stored in organic material. Nitrogen is accumulated through fixation, and both phosphorous and potassium are derived from minerals. Availability of nutrients depends on total content, rate of release (low solubility of inorganic phosphorous and the micro-nutrient cations keeps them available to plants), and potential for leaching.
- *Acid Soils.* Where soils have excess acidity, plants are susceptible to the toxic effects from elements such as aluminum and manganese. Furthermore, acidity limits the availability of certain nutrients and interferes with important biological processes. Acidity may be reduced by liming the ground with crushed limestone or products derived from it. The lime supplies hydroxyl ions that neutralize the acids and raise the pH of the soils. The reaction time can be reduced by using finely ground lime and mixing it with the soil.
- *Saline Soils.* Where soils have excess soluble salts, these salts compete with plants for available water. The situation normally occurs in dry climates where drainage is inadequate. If the salts are compounds of sodium, they may contribute towards the accumulation of exchangeable sodium, and sodic soils are caused by excess exchangeable sodium. Sodic soils tend to have poor tilth because of the lack of structure and they may also have a high pH which may restrict the access of nutrients for plants. The best way to improve sodic soils is reduction by leaching with amendments containing soluble calcium which will displace exchangeable sodium.

Classifications of Soil Horizons

Three kinds of symbols are used in various combinations to designate horizons and layers. These are capital letters, lower case letters, and Arabic numerals. Capital letters are used to designate the master horizons and layers; lower case letters are used as suffixes to indicate specific characteristics of master horizons and layers; and Arabic numerals are used both as suffixes to indicate vertical subdivisions within a horizon or layer and as prefixes to indicate discontinuities. Soil profiles consist of the master horizons and layers O, A, E, B, C and R; each of these is subdivided into categories that are distinguished by important properties. Most horizons and layers are given a single capital letter symbol; however, some require two. The classification of horizons is set forth below and was adopted by the Soil Survey in October 1981 and revised in 1993 by the Soil Survey Division Staff.

Horizon and Layer Descriptions

The horizon and layer descriptions taken directly from the Soil Survey Division Staff (1993) unless otherwise noted:

- *O horizons or layers*: Layers dominated by organic material. Some are saturated with water for long periods or were once saturated but are now artificially drained; others have never been saturated.

Some O layers consist of undecomposed or partially decomposed litter, such as leaves, needles, twigs, moss, and lichens, that has been deposited on the surface; they may be on top of either mineral or organic soils. The mineral fraction of such material is only a small percentage of the volume of the material and generally is much less than half of the weight. Some soils consist entirely of material designated as O horizons or layers. An O layer may be on the surface of a mineral soil or at any depth beneath the surface, if it is buried.

- *A horizons*: Mineral horizons that formed at the surface or below an O horizon, that exhibit obliteration of all or much of the original rock structure, and that show one or more of the following: (1) an accumulation of humified organic matter intimately mixed with the mineral fraction and not dominated by properties characteristic of E or B horizons or (2) properties resulting from cultivation, pasturing, or similar kinds of disturbance.
- *E horizons*: Mineral horizons in which the main feature is loss of silicate clay, iron, aluminum, or some combination of these, leaving a concentration of sand and silt particles. These horizons exhibit obliteration of all or much of the original rock structure.

An E horizon is usually, but not necessarily, lighter in color than an underlying B horizon. In some soils the color is that of the sand and silt particles, but in many soils coatings of iron oxides or other compounds mask the color of the primary particles. An E horizon is most commonly differentiated from an overlying A horizon by its lighter color. It generally has less organic matter than the A horizon.

The E horizon typically has a light gray or whitish color because nearly all the iron and organic matter has been removed. It is present only in areas of relatively high precipitation. It generally occurs immediately beneath an O or an A horizon or immediately above a very slowly permeable horizon.

- *B horizons*: Horizons that formed below an A, E, or O horizon and are dominated by obliteration of all or much of the following: (1) illuvial concentration of silicate clay, iron, aluminum, humus, carbonates, gypsum, or silica, alone or in combination; (2) evidence of removal of carbonates; (3) residual concentration of sesquioxides; (4) coatings of sesquioxides that make the horizon conspicuously lower in value, higher in chroma, or redder in hue than overlying and underlying horizons without apparent illuviation of iron; (5) alteration that forms silicate clay or liberates oxides or both and that forms granular blocky, or prismatic structure if volume changes accompany changes in moisture content; or (6) brittleness.

In some soils, the B horizon has the brightest yellowish-brown or reddish-brown color. In others, it has the most pronounced blocky or prismatic structure. Many B horizons have more clay than any other horizons in the profile and show evidence of clay skins.

- *C horizons*: Horizons or layers, excluding hard bedrock, that are little affected by pedogenic processes and lack properties of O, A, E, or B horizons. The material of C layers may be either like or unlike that from which the solum presumably formed. The C horizon may have been modified even if there is no evidence of pedogenesis.

Included as C layers are sediment, saprolite, unconsolidated bedrock, and other geologic materials that are commonly uncemented and exhibit low or moderate excavation difficulty. Layers that have accumulations of silica, carbonates, or gypsum or more soluble salts are included in C horizons, even if indurated.

Any material that can be dug with a shovel but has not changed appreciably by soil forming processes is considered a C horizon.

- *R layers*: Hard Bedrock. Granite, basalt, quartzite, sandstone and limestone are examples of bedrock.

Soil Texture

Texture is the relative proportion of different particle sizes, similar to the property of sorting used by geologists. Geologists, engineers, and soil scientists all use separate systems of particle size classification, although the size categories are very similar. The term fine earth refers to the finer sizes of the soil which are smaller than 2 mm in diameter. Material more than 2 mm in diameter includes rock fragments which are broken down into pebbles, cobbles, stones, and boulders, depending on diameter.

Particle-Size Classification by the USDA of Material < 2 mm in Diameter:

Size Fraction (Coarse Fragments) Diameter	
Boulder	Above 600 mm
Stones	250 - 600 mm
Cobbles	75 - 250 mm
Gravels	2 - 75 mm

The USDA uses the following size separates for the <2 mm material (Soil Survey Division, 1993):

Soil Separates	
Very Coarse Sand	2.0-1.0 mm
Coarse Sand	1.0-0.5 mm
Medium Sand	0.5-0.25 mm
Fine Sand	0.25-0.10
Very Fine Sand	0.10-0.05 mm
Silt	0.05-0.002 mm
Clay	<0.002 mm

To be called soil, the USDA says a deposit of mineral matter should contain at least 10 percent sand, silt or clay. The large sizes impair the ability of the soil to provide water and nutrients to plants and present obstacles to cultivation.

- *Particle Shape and Size.* Both particle shape and particle size determine the pore

spaces and permeability of the soil. Particle shape also determines the cohesive and adhesive properties. These properties express the force with which the mass of soil clings to itself (cohesion) or to other objects (adhesion). Flat surfaces in parallel have the strongest contact. These properties are measured by the firmness of clods when dry or the plasticity when wet.

- *Sand Properties.* Sand grains are large enough to be seen with the unaided eye and felt individually when rubbed between fingers. Sand grains, because of size and roughness, make only minimum contact with other grains. Also, sand is not sticky, does not form stable aggregates, and is permeable to air and water.
- *Silt.* Individual particles of silt cannot be seen or felt between the fingers. Silt is more cohesive and adhesive than sand, but still contributes little to stable aggregate formation in soil.
- *Clay.* Clay is so small that the individual particles can only be seen under an electron microscope. Although clay is typically sticky and plastic when wet, some clay minerals express these qualities better than others. Expanding clays are characterized by their ability to swell when wet and shrink when dry, commonly forming large cracks.

Soil Texture is the weight proportion of the separates for particles less than 2 mm as determined from a laboratory particle-size distribution. The field texture, which is determined by feel, is referred to as apparent because it is not an estimate of the results of a laboratory operation. Sand particles feel gritty and can be seen individually without magnification; they have a smooth feel to the fingers when dry or wet. In some places, clay soils are sticky; and in others they are not. Where soils are dominated by montmorillonite clays, they feel different from soils that contain similar amounts of micaceous or

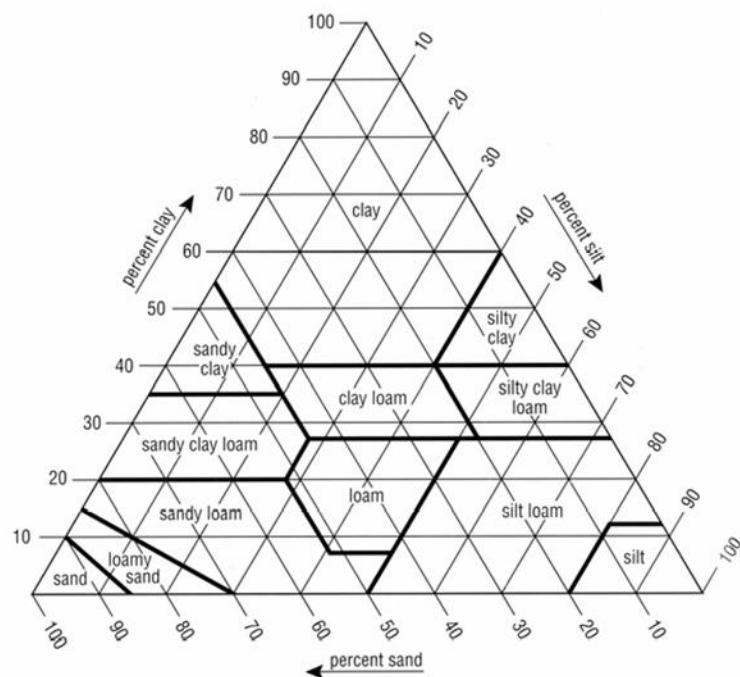


Chart showing the percentages of clay, silt, and sand in the basic textural classes.

kaolinitic clay.

The texture classes are sand, loamy sands, sandy loams, loam, silt loam, silt, sandy clay loam, clay loam, silty clay loam, sandy clay, silty clay, and clay. Subclasses of sand are subdivided into coarse sand, sand, fine sand, and very fine sand. Subclasses of loamy sands and sandy loams that are based on sand size are named similarly.

Field Determination of Texture

The particle size distribution of material less than 2 mm in diameter is determined in the field by feel. The distribution of mineral particles of a soil is an important property that influences many other properties. In the field the various particle size classes can be estimated by moistening the soil to maximum plasticity and working it with the fingers to eliminate any aggregation effects. It is possible to determine the relative proportions of sand, silt, and clay in soil in the field almost as accurately as can be accomplished in a laboratory. The sand fractions are gritty, the silt fractions are smooth, and the clay fractions are plastic and sticky.

Soil Structure

Structure is the shape formed when individual particles cluster together into aggregates called peds. Soil structure refers to units composed of primary particles where the cohesion within units is greater than the adhesion among units. Under stress, the soil mass tends to rupture along predetermined planes or zones which form the boundary of the units. The term “structural unit” applies to any repetitive soil body that is commonly bounded by planes or zones of weakness that are not an apparent result of compositional differences. A structural unit that is the consequence of soil development is called a ped. The surfaces of peds persist through cycles of wetting and drying in place.

Soils devoid of structure are referred to as structureless. In structureless layers or horizons, no units are visible in place after the soil has been gently disturbed, such as by tapping a spade containing a slice of soil against a hard surface or dropping a large fragment on the ground. When structureless soils are ruptured, soil fragments, single grains, or both result. In soils that have structure, the units are described in terms of shape, size, and grade (distinctness).

Soil structure refers to the way that primary soil particles aggregate into specific structural units called peds. Therefore, when the soil is wet and the peds are swollen, the aggregates fit tightly together. Prismatic structure may be caused by slow, prolonged shrinkage of the soil, with cracks formed perpendicular to the drying surface. Blocky structures may be caused from rapid shrinkage of clay-rich soil while drying. Platy structures may be caused by very rapid drying with formation of cracks parallel to the surface of drying.

Certain soil structures are typical for different parts of the soil profile. For example, surface soil

is generally characterized by granular structures. Structure formation is normally less developed with depth and the peds tend to be the larger prisms or blocky shapes. Structureless horizons generally form below the zone disturbed by fluctuating moisture conditions or roots.

Shape

Five basic shapes of structural units are recognized in soils:

- *Platy*: The units are flat and platelike. They are generally oriented horizontally. Lenticular platy structure has a shape that is thickest in the middle and thins towards the edges.
- *Prismatic*: The individual units are bounded by flat to rounded vertical faces. Units are distinctly longer vertically, and the faces are typically casts or molds of adjoining units. Vertices are angular or sub-rounded; the tops of the prism are somewhat indistinct and normally flat.
- *Columnar*: The units are similar to prisms and are bounded by flat or slightly rounded vertical faces. The tops of columns, in contrast to those of prisms, are very distinct and normally rounded.
- *Blocky*: The units are blocklike or polyhedral. They are bounded by flat or slightly rounded surfaces that are casts of the faces of surrounding peds. Typically, blocky structural units are nearly equidimensional but grade to prisms and to plates.
- *Granular*: The units are approximately spherical or polyhedral and are bounded by curved or very irregular faces that are not casts of adjoining peds.

Soil Consistency

The consistency of a soil depends on the degree and kind of cohesion and adhesion, or from a negative sense, the resistance to deformation or rupture. The evaluation of soil consistency relies on three moisture contents which include (1) consistency when wet is determined at slightly above field capacity; soil material is pressed between the thumb and finger and its degree of stickiness is determined; (2) consistency when moist is determined at a moisture content between air dry and field capacity; consistency is characterized by tendency to break into smaller particles rather than into powder; it has the ability to cohere again if pressed together; and (3) the consistency of dry soil materials is characterized by brittleness, rigidity and a tendency to deform to a powder or to fragments with sharp edges; it also lacks the ability to cohere if pressed together again.

Soil Porosity

Soil porosity depends on the shape, size, and abundance of cracks, passages and other soil

cavities. Six criteria may be used to describe soil pores: size, abundance, shape, distribution, continuity, and orientation; however, in field descriptions, porosity is expressed by abundance, size and shape. Pore sizes are described by the following diameter classes: micro (less than 0.075 mm), very fine (0.075-1 mm), fine (1-2 mm), medium (2-5 mm), and coarse (over 5 mm). Soils with a pore diameter of more than 0.075 mm will generally drain freely. Interstitial soil pores are strongly influenced by soil moisture. In the case where peds fit together, the interstitial pores are larger when the soil is dry, but when the peds swell because of increasing moisture, the interstitial pore volume may be reduced very significantly. Where peds are spheroidal and do not fit together, the size of the interstitial pores is less dependent on the moisture content of the soil.

Soil Color

Soil color is an important characteristic for the classification and study of soil, and is useful for soil identification when combined with other soil characteristics such as structure. Humus, the most stable component of organic matter, is the most common coloring agent in topsoil. Soil color can indicate the extent of mineral weathering, the existence of organic material and the state of aeration. The red, yellow, and brown colors are caused by the secondary iron oxides and the black oxides are manganese oxides. Organic material darkens soils, particularly under grasslands where a few percent organic material may turn the soil very dark. In water-saturated soils with poor aeration, iron and manganese assume reduced forms and cause the soil to have the grayish or bluish colorations of gleying. Light-colored layers may be caused from leaching of iron minerals from the mineral matrix.

The parent materials are important contributors to soil color, particularly in young soils where soils tend to have the color of the parent rock. Mafic rocks, which are rich in iron, commonly produce red or brown soils.

Mottling in soils may be recorded by describing the color of the matrix and the colors of the principal mottles as well as the mottling pattern. Mottles, which may be gray, reddish brown, and yellow, are typically present in poorly drained soils where the water table fluctuates. The amount of gray tends to increase in wetter organic soils. Mottle patterns should be described according to their contrast, size and abundance.

Soil Classification

Soil classification is the systematic organization of soils into groups or categories on the basis of their properties and characteristics. Subdivisions of these groups are accomplished on the basis of specific property differences. There are six categories in the United States:

Orders (10)

Suborders (29)

Great Groups

Subgroups
Families
Soil Series

Soil Sampling and the Pedon

Soil properties cannot be determined from sampling the surface. Soil is a three dimensional body that requires the study of its horizons or layers. Sampling of soils is complicated by the fact that a soil is typically is variable, both vertically and horizontally. The concept of a pedon was developed to make possible a clear basis for soil descriptions and the selection of soil samples. Where describing and sampling soil, the pedon has the smallest volume possible for representing the nature and arrangement of soil horizons and variability in the properties that are preserved in samples. Like a unit cell, the pedon has three dimensions with a minimal horizontal area arbitrarily set at 1m² but ranges to 10 m², depending on soil variability. The lower limit of a pedon is the typically transitional interface between soil and nonsoil. It is important that the lateral dimensions are sufficiently large to represent the nature of the horizons and their variability. The polypedon may be defined as a set of contiguous pedons.

Surface Water

Running water is the most powerful agent of erosion. Continents are eroded primarily by running water at an average rate of 1 inch every 750 years. The velocity of a stream increases as its gradient increases, but velocity is also influenced by factors such as degree of turbulence, position within the river, the course of the stream, the shape of the channel and the stream load.

River Cycles

Stages in the cycle of river erosion are labeled as youth, maturity and old age. Each stage has certain characteristics that are not necessarily related to age in years—only phases in development. Typically, rivers tend to have old-age-type development at their mouths and youthful development at their upper reaches. So the three stages may grade imperceptibly from one to another and also from one end of the stream to the other.

The youthful stage is characterized by rapid down cutting, high stream gradient, steep-sided valleys with narrow bottoms and waterfalls. The mature stage is characterized by a longer, smoother profile and no waterfalls or rapids.

Gradient is normally expressed as the number of feet a stream descends each mile of flow. In general, a stream's gradient decreases from its headwaters toward its mouth, resulting in a longitudinal profile concave towards the sky.

Base Level

The base level of a stream is defined as the lowest level to which a stream can erode its channel. An obstacle such as a resistant rock across a stream can create a temporary base level. For example, if a stream passes into a lake, it cannot erode below the level of the lake until the lake is destroyed. Therefore different stretches of a river may be influenced by several temporary base levels. Of course the erosive power of a stream is always influenced by the ocean which is the ultimate base level below which no stream can erode with the exception of a few areas below sea level. Many streams in Idaho eventually reach the ocean through the Columbia River.

Transportation of Material

Running water transports material in 3 ways: solution, suspension and by rolling and bouncing on the stream bottom. Dissolved material is carried in suspension. About 270 million tons of dissolved material is delivered yearly to the oceans from streams in the United States. Particles of clay, silt and sand are generally carried along in the turbulent current of a stream. Some particles are too large and heavy to be picked up by water currents, but may be pushed and shoved along the stream bed.

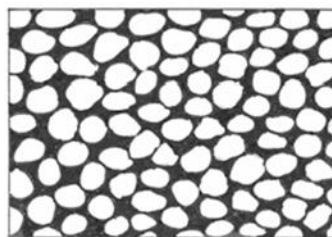
Waterfalls

Waterfalls are a fascinating and relatively rare occurrence. Waterfalls may be caused in several ways. For example, where a relatively resistant bed of rock overlies less resistant rock, undermining of the less resistant rocks can cause a falls. Waterfalls are temporary features in the history of a stream as they are created by a temporary base level. As time passes, falls may slowly retreat upstream, perhaps as rapidly as several feet per year. There are many spectacular waterfalls in Idaho, including the 212-foot-high Shoshone Falls in the Snake River Canyon just north of Twin Falls.

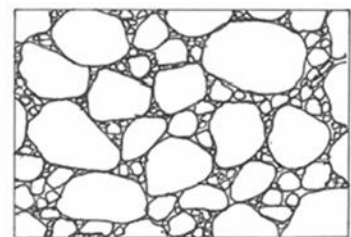
Ground Water

Ground water is the water that lies below the surface of the ground and fills the pore space as well as cracks and other openings.

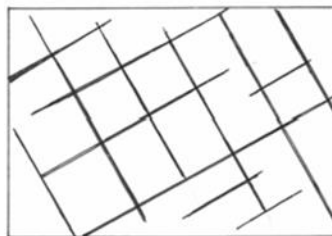
Porosity is the percentage of a rock's volume that is taken up by openings. Most sedimentary rocks such as sandstone, shale and limestone can hold a large percentage of water. Loose sand may have a porosity of up to 40 percent; however, this may be



High porosity in well-sorted sedimentary rock



Low porosity in poorly sorted sedimentary rock



Porosity in rocks



Porosity from solution

The porosity of a rock is dependent on the size, shape, and arrangement of the material composing the rock. The top two drawings represent primary porosity and the lower two represent secondary porosity.

reduced by half as a result of recrystallization and cementation. Even though a rock has high porosity, water may not be able to pass through it. Permeability is the capacity of a rock to transmit a fluid such as water. For a rock to be permeable, the openings must be interconnected. Rocks such as sandstone and conglomerate have a high porosity because they have the capacity to hold much water.

Water Table

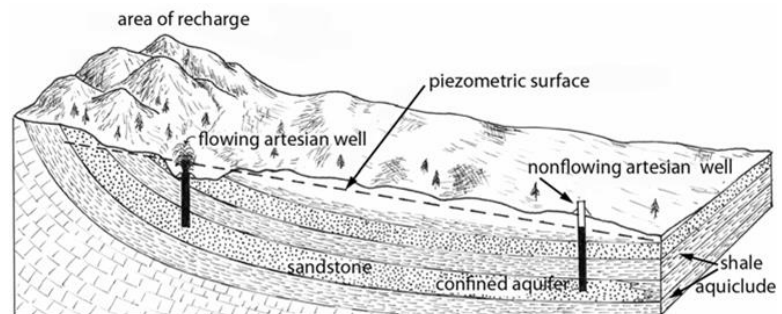
In response to gravity, water seeps into the ground and moves downward until the rock is no longer permeable. The subsurface zone in which all openings of the rock are filled with water is called the zone of saturation. The upper surface of the zone of saturation is called the water table. The zone that exists between the water table and the ground surface is called the zone of aeration. In order to be successful, a well must be drilled into the zone of saturation. The velocity at which water flows underground depends on the permeability of the rock or how large and well-connected the openings are.

Springs occur where water flows naturally from rock onto the surface of the land. Springs may seep from places where the water table intersects the land surface. Water may also flow out of the ground along fractures.

Aquifers

An aquifer is a body of saturated rock through which water can easily move. Aquifers must be both permeable and porous and include such rock types as sandstone, conglomerate, fractured limestone and unconsolidated sand and gravel. Fractured volcanic rocks such as columnar basalts also make good aquifers. The rubble zones between volcanic flows are generally both porous and permeable and make excellent

aquifers. In order for a well to be productive, it must be drilled into an aquifer. Rocks such as granite and schist are generally poor aquifers because they have a very low porosity. However, if these rocks are highly fractured, they make good aquifers. A well is a hole drilled into the ground to penetrate an aquifer. Normally such water must be pumped to the surface.



An artesian water system has several requirements: (1) the aquifer must be exposed at the ground surface, (2) there must be an adequate amount of precipitation to maintain a full aquifer, and (3) the aquifer must be confined above and below by aquicludes. The sloping dashed line represents the piezometric surface or the artesian-pressure surface which is the highest level water can rise in an artesian system. If the wellhead is below the piezometric surface, then the well will be free flowing.

If water is pumped from a well faster than it is replenished, the water table is lowered and the

well may go dry. When water is pumped from a well, the water table is generally lowered into a cone of depression at the well. Ground water normally flows down the slope of the water table towards the well.

Snake River Plains Aquifer

- *Size.* The Snake River Plain aquifer underlies most of the Snake River Plain or about 10,000 square miles. The upper surface of the aquifer (water table) ranges from about 200 feet to about 1,000 feet below the ground surface; however, in most places the aquifer is no deeper than 450 feet. In the eastern Snake River Plain the water table slopes to the southwest at gradients ranging from 1 m/km to 6 m/km. Consequently, the gravity and weight of the water in the basalt layers north of the river drives the water emerging from Thousand Springs. The aquifer steepens at the west end of the eastern plain, possible because the underlying basalt aquifer thins.

The aquifer is immense and contains vast quantities of water. The storage capacity may be equal to several hundred times the annual flow of the Snake River.

Approximately 8 million acre-feet of water are moved in and out of the aquifer every year.

- *Composition.* The aquifer consists primarily of flat-lying Pleistocene basalt flows with interbedded sedimentary rocks of both fluvial and lacustrine sources. Rhyolite, which has low permeability, underlies the basalt and forms most of the base of the aquifer throughout the plain. The rhyolite may be much less permeable than the basalt because fractures tend to be filled by chemical precipitates. Basalt lava flows have permeable rubble and vesicular or breccia zones at both the tops and bottoms of flows. Water can also move vertically through vertical cooling fractures in the basalt flows. Sedimentary interbeds consisting of sand, silt, clay and ash also offer permeable strata between basalt flows.

The aquifer is a water-saturated zone that overlies impermeable rocks such as the silicic volcanics. The upper surface of the aquifer (water table) varies substantially on the recharge rate from precipitation and irrigation.

- *Source of the Water.* The aquifer is fed by precipitation from drainage areas in the northern mountains covering approximately 35,000 square miles. The water running off the large drainage basins such as the Big and Little Lost River and Birch Creek, all sink into the permeable lavas at the northeast end of the plain and enter the aquifer. Only the Big Wood River and the Henry's Fork of the Snake manage to cross the northern Snake River Plain and make it to the Snake River; all others seep entirely into the lava and disappear into the aquifer. Approximately 3 million acres of

agricultural land on the Snake River Plain are irrigated, of which 1 million acres are irrigated from wells tapping the aquifer.

- *Pollution of the Aquifer.* Pollution of the Snake River Plain aquifer by localized sources such as fertilizer, potato processing plants, feedlots and the Idaho National Engineering Laboratory (INEL) is a problem in certain areas but does not at this time present a serious overall threat to the aquifer. These pollution sources are localized and because of the immense size of the aquifer, depth to the water table and lack of precipitation over the aquifer, the pollution that enters the aquifer is diluted.

Approximately 15 million cubic feet of radioactive waste is stored at the INEL over the aquifer. An elaborate monitoring system has been established and many detailed studies have been made of the geologic and hydrologic characteristics of the aquifer and the related rocks and structures. According to data from monitoring wells, there has not been significant deterioration of the Snake river Plain aquifer from storage of radioactive wastes. Samples of water from the Thousand Springs area indicate only natural background levels of radiation.

Thousand Springs

The Thousands Springs area is situated on the north side of the Snake River Canyon between Bliss and Twin Falls. Between Milner Dam and King Hill, the Snake River has cut a canyon several hundred feet deep for a distance of almost 90 miles. The north wall of the canyon intercepts the Snake River Plain aquifer where many spectacular springs pour out of the truncated pillow basalts and other permeable zones. This area embraces 11 of the 65 springs in the United States that produce more than 72,400 acre-feet of water annually. Unfortunately, many of the most spectacular springs are on private lands and are conveyed through flumes and pipelines rather than gushing out of the aquifer and down the canyon wall. Until the early part of this century when the water was developed for power and irrigation, the springs were a spectacular scenic feature with springs pouring from outlets along the canyon forming an almost continuous wall of water over the north side of the canyon for hundreds of yards.



Large springs at Thousand Springs, Idaho. The water discharges from pillow basalts and flows down over the edge of a flow on the north wall of the Snake River Canyon. The height of the falls is approximately 180 feet. When this photo was taken by I.C. Russell in 1901, the springs poured from outlets along the canyon wall forming a continuous wall of water for hundreds of meters. This view is no longer in existence because much of the water is diverted for commercial uses as it emerges from the springs.

The aquifer is a major source for the Snake River. Between April and October, most of the Snake River water above twin Falls is diverted for irrigation. Consequently, almost all the water in the Snake River downstream from Twin Falls originates from springs emerging from the Snake River Plain aquifer.

The origin of Thousand Springs occurred from a unique sequence of geologic events and processes. Periodically, lava flows originating from sources on the north side of the Snake River canyon flowed south and dammed the canyon forming a large lake. Once the lake was formed, lava continued to flow into the lake. When basalt lava flows into a lake, the lava tends to separate into discrete ellipsoidal pillow lavas that are highly shattered or fractured from rapid cooling.

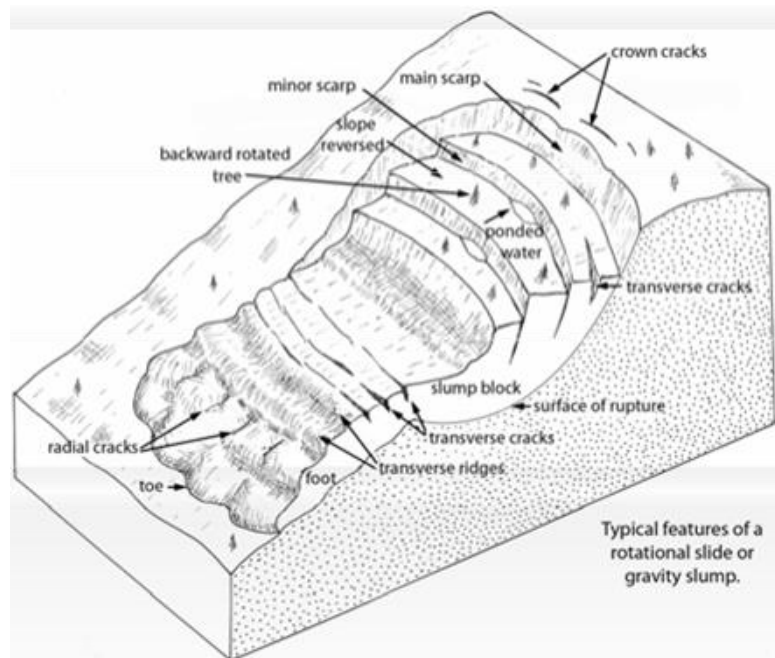
These pillow lavas have a very high porosity and permeability. However, the first lava which entered the canyon and formed the dam is neither very permeable nor porous so it confines the flow of water to the pillow basalt aquifer on its west end.

When the canyon was filled by pillow basalts, the Snake River was forced to the south to cut a new channel in the older Tertiary basalts and sediments where the old pillow-basalt-filled channel is exposed at the north wall of the canyon. The springs must pass over the upper edge of the Tertiary basalt which acts as a sill by maintaining the water table to the top of the lowest edge of the Tertiary basalt.

Mass Movement

Mass movement is the movement of surface material caused by gravity. Landslides and rockfalls are examples of very sudden movements of this type. Of course, geological agents such as water, wind, and ice all work with gravity to cause a leveling of land.

Water aids in the down-slope movement of surface material in several ways. Water adds weight to the soil; it fills pore spaces of slope material and it exerts pressure which tends to push apart individual grains. This decreases



the resistance of the material to movement. Landslide is a general term that is commonly broken down into the more specialized terms such as slump, rockslide, debris slide, mudflow, and earthflow.

Slump

A slump is a downward and outward movement of rock or unconsolidated material moving as a unit or series of units. Large blocks of material move suddenly downward and outward along a curved plane.

Rockslide

Rockslides are the most catastrophic type of landslide. They involve a sudden rapid slide of bedrock along planes of weakness. Rockslides are very common in the over-steepened canyons and drainages of Idaho, particularly in those areas like the Salmon River Canyon where more than 5,000 feet of elevation may exist between the ridge tops and the canyon bottoms.

Debris Flow

A debris flow is a mass of transported saturated rock particles of all sizes. This type of landslide is caused by a sudden flood of water from a cloudburst in semi-arid country or a sudden thaw. The flood waters carry the soil and rocks from a large slope area and transport them down a gulch or canyon. Then the water and debris move down the canyon and spread out as an alluvial fan on the gentle slopes below. Mudflows are very common in the semi-arid areas of southwestern Idaho as well as high-relief areas of central Idaho.

Talus

A talus slope is developed by an accumulation of rock fragments at the foot of a cliff or ridge. Rock fragments break loose from the cliff above, roll down the slope and pile up in a heap of rock rubble. Individual talus forms as a half-cone with the apex pointing upwards. In most cases a series of half cones coalesce around the base of a mountain.

Horseshoe Bend Hill Slide Area

Landslides are a very common occurrence on the Horseshoe Bend Hill area between Boise and Horseshoe Bend. From the highway you can easily see earthflows of less than one year old as well as those much older. The more recent flows show fresh brown crescentric cracks where the fresh Earth is exposed. The older flows are more difficult to identify because vegetation has grown over the scarp areas. Most of the slides occur during the spring when the ground surface is saturated with water. Placement of this major north-south highway over an active slide area has resulted in a section of highway constantly deformed and broken by the slowly-moving land surface.

Warm Springs Mesa Slide

Warm Springs Mesa is situated immediately south of Table Rock in east Boise. The entire Warm Springs Mesa is a 300-acre landslide. The construction of Warm Springs Avenue along the toe of the landslide has caused an over-steepened natural slope. There has been sliding along this over-steepened slope for years and debris is constantly falling on Warm Springs Avenue. Although geologists have long cautioned against development until study of the effect of increased water in the sediment is completed, development of the subdivision has not stopped.

A number of investigators have determined that the combination of the over-steepened slopes coupled with ground water causes the sliding. In the past, the additional ground water derived from the new residential uses is also believed to have had an adverse impact on the sliding activity.

The Warm Springs Mesa slide originated in an area adjacent to Table Rock. Perhaps an earthquake suddenly dislodged the material and caused a sudden movement of a large earth mass down slope in a southwest direction some 1,200 feet towards the Boise River. The surface of the landslide is now re-vegetated but has the typical hummocky rolling topography of a landslide area. Numerous large boulders of sandstone are exposed chaotically over the surface but are particularly abundant on the over-steepened south slope. The large sandstone boulders are derived from the sedimentary rock that is now exposed at Table Rock. From an airplane perspective, one can readily envision both the source and the total extent of the fan-shaped slide deposit.

Bliss Landslide

On July 24, 1993, a landslide affecting an area about 100 acres occurred about 0.6 miles south of the town of Bliss Idaho. According to eye-witness accounts, the toe of the slide temporarily blocked the Snake River and forced the river to cut a new channel south of the slide. The movement over a 1-month period also wiped out almost 0.5 miles of the paved Shoestring Road. The slide occurred in Pliocene sediments of the Glens Ferry Formation which overlies pillow lavas of Tertiary Banbury Basalt. Glens Ferry



Vertical aerial photography of the Bliss landslide near Bliss, Idaho. The landslide initially completely dammed the Snake River, but within hours eroded a new channel through the landslide debris.

sediments that make up the disrupted slide material are lacustrine clays of the Yahoo Clay. The Yahoo Clay was deposited when the McKinney Basalt dammed up the Snake River and formed a lake approximately 70,000 years ago.

Outstanding textbook examples of most of the features typically associated with landslides can easily be observed at the Bliss Landslide. These features include rotated slump blocks, hummocky surface at the earthflow portion near the toe, tension cracks, grabens, water ponds, staircase topography and a large head scarp. The passage of time has subdued many of these classic features and made it more difficult to recognize them.

Other Idaho Landslides

Landslides are a common sight in the mountainous areas of Idaho. Once you know what to look for, they can be readily identified by the presence of a rupture in the vegetative cover exposing fresh earth or by a hummocky lower surface.

Caves in Idaho

Caves in Idaho generally fall into one of three types: corrosion caves, solution caves and lava caves. Corrosion caves are formed by erosive action of water, waves or currents on a relatively soft rock. These caves generally occur at the edge of a river or lake.

Corrosion Caves

Corrosion caves are generally shallow and not as impressive as lava caves or solution caves. Archeologists, however, have found that early Americans commonly camped in small corrosion caves while hunting or fishing in the vicinity of water bodies. Rock shelters are also formed by erosion recessing the lower rocks in a cliff and leaving an overhanging rock shelter. Many of these rock shelters have yielded valuable information on the culture and migration patterns of early Idahoans.

Solution Caves

Solution caves are formed by slightly acidic ground water circulating through fractures in limestone. This water is capable of dissolving great quantities of solid rock. As time passes, the openings become larger and larger until they may be large enough for a man to pass through. Cone-shaped forms called stalactites are deposited by underground water. Stalactites are composed of calcium carbonate and look like icicles hanging from cave roofs. Stalagmites are similar in composition and origin to stalactites but are formed from ground water dripping on the cave floor. The best examples of solution caves are found in the Paleozoic carbonate rocks in eastern Idaho.

Lava Caves

Lava caves, also known as lava tubes, form in the central portion of a lava flow. Immediately after the flow is extruded, the outer margins of the flow cool and freeze in place, including the bottom, sides and top. Although the outer margin of the flow has solidified into basaltic rock, the central core is still molten and continues to flow towards the flow front. When the source of lava is cut off, the lava flows out the end of the tube and leaves a cave. These caves are typically 10 to 20 feet in diameter. They are characterized by both stalactites and stalagmites formed by lava dripping off the roof of the tube. Basalt flows on the Snake River Plain have many excellent examples of these caves. Many such caves are found where a thin portion of the roof collapses and leaves a precarious entrance to the cave.

Glaciation

Ice Age

The term ice age is used here to refer to a period of relatively cold temperatures, throughout Earth history. These cold periods may last anywhere from a few million years to tens of millions of years. Typically, these ice ages are characterized by cold glacial stages alternating with warmer interglacial stages. Glacial stages within an ice age generally have a duration measured in thousands or tens of thousands of years. The Pleistocene ice age has run for the last 2 to 3 million years and the last glacial stage had its greatest expansion 21,000 to 17,000 years ago, but started 115,000 years ago and ended about 10,000 years ago.

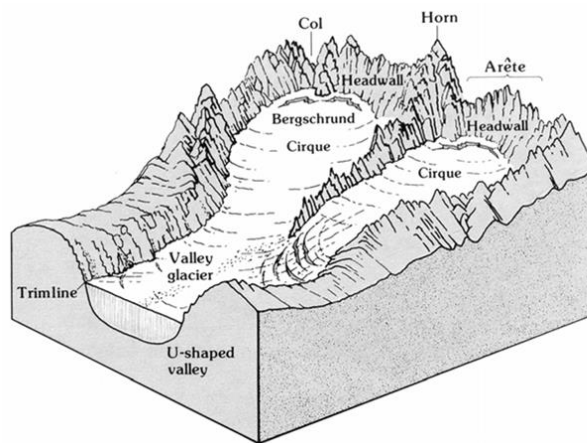


Illustration shows erosional features sculpted by glaciers (from figure 56, Huber, 1975).

The Pleistocene ice age had a large role in shaping the landforms in the Boise valley and the mountains to the northeast and southwest of the plain. Evidence of this ice age is now widespread throughout the high country of Idaho. Most of the high country of Idaho, including areas over 5000 feet in elevation, has been modified by alpine glaciation.

Climate and Glaciation. The Pleistocene climate was probably semiarid like the present climate; however, as the mountains were glaciated, the western Snake River Plain was cooler with longer periods of snow accumulation, and large runoff during the spring snowmelt. Patterned ground is common on the higher terraces and uplands in the Boise Valley. These features are thought to be caused by intense freeze and thaw cycles.

A glacier is a large, slow-moving mass of land ice that moves under its own weight. It is formed by the accumulation, compaction and recrystallization of snow. For a glacier to form, more snow must accumulate than is melted. Two types of glaciation are recognized and both have affected Idaho. Alpine glaciation of smaller areal extent is found in mountainous regions. Continental glaciation has covered a large part of the continent with a huge ice sheet. Both types of glaciation have dramatically changed the landscape.

Glacial Erosion

Slow-moving glaciers plowed up soil and loose rock and plucked boulders from outcrops. This material, incorporated in the glacier, was used as an abrasive to grind down, polish and scratch the exposed outcrops in its downward path. In this way glaciers soften landscapes by wearing down hilltops and filling in valleys. In mountainous areas, glaciers confined to valleys (valley glaciers) scooped out and widened the valleys leaving a U-shaped cross profile. Stream erosion normally leaves a V-shaped valley so that the presence of a U-shaped valley is strong evidence that the valley was shaped by a glacier.

A glacial cirque is a steep-sided, rounded, bowl-shaped feature carved into a mountain at the head of a glacial valley. In the cirque, snow accumulates and eventually converts to glacier ice before heading down the glacial valley. A horn is the sharp peak that remains after cirques have cut back into a mountain on several sides. Sharp ridges called *arêtes* separate adjacent glacially-carved valleys. The Sawtooth Mountains of Idaho offer exceptional examples of glacial erosional features such as U-shaped valleys, cirques, horns, and *arêtes* as well as smaller features such as polished and striated bedrock.



A tarn lake and its associated U-shaped valley; a tarn lake occurs where a glacier gouges out a basin in the bedrock as it moved down the valley. Tarn lakes tend to be small in size and occur in mountainous terrain. Central Idaho.

Glacial Deposition

As glaciers move down valley, rock fragments are scraped and plucked from the underlying bedrock and the canyon walls. Most of these rock fragments are angular. When the material picked up and transported by the glacier is deposited, it is called till. Glacial till consists of unsorted fragments ranging from clay size to boulder size, all mixed together with no layering.

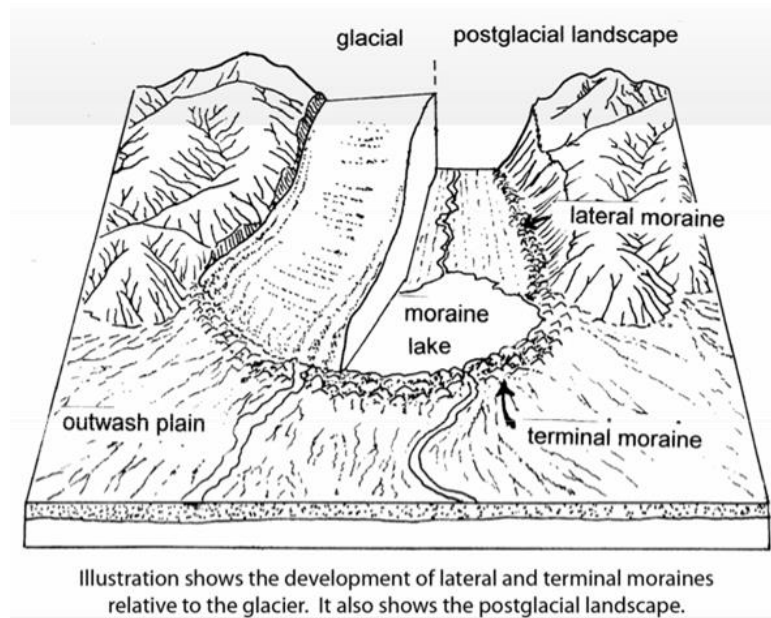
Glaciers can easily carry any size rock fragment including boulders as large as a house. An erratic is a huge, ice-transported boulder that is not related to bedrock in the vicinity.

A moraine is a body of till deposited by a glacier. Ridge-like piles of ice left at the sides of a glacier are called lateral moraines. Medial moraines are developed where two glaciers come together and their lateral moraines merge and continue down glacier as a single long ridge of till. An end moraine is piled up along the front edge of ice at the down-slope terminus of a glacier. Valley glaciers tend to leave an end moraine with the shape of a crescent ridge. Redfish Lake near Stanley is a glacial lake. The lake occupies a U-shaped, glacially-carved valley and the water is contained on the sides by lateral moraines and on the north end near the lodge by a terminal moraine.

Streams that drain glaciers are heavily loaded with sediment, especially during the summer months. These outwash streams form a braided pattern and their deposits are layered as are all stream deposits. Thus, outwash deposits can be distinguished from till which is unsorted.

Pleistocene Ice Age in Idaho

Glaciation began in the northern hemisphere more than two million years ago. However it was not until about 100,000 years ago that glaciers formed in southern British Columbia began moving southward following major south-trending valleys.



Poorly sorted glacial till exposed in a road cut near Stanley, Idaho.

Central and northern Idaho were exposed to glaciation during Early and Late Pleistocene time. Northern Idaho was first covered by a continental ice sheet, and later carved by mountain glaciation. Evidence of Pleistocene glaciation can be seen in mountainous areas at elevations as low as 5,000 feet above sea level.

In northern Idaho, the continental ice sheet moved from the Canadian ice fields towards the south into northern Idaho. This ice sheet probably extended no further south than the north end of Coeur d'Alene Lake. The continental ice sheet, originating in the Canadian ice fields, invaded northern Idaho repeatedly. Slow advances were followed by retreat as the climate warmed or cooled. During the melting phases, deposits of sand and gravel accumulated at the margins of the ice lobe. These deposits are commonly called recessional moraines. The grinding of the moving ice sheet left scratched, grooved and polished surfaces on much of the bedrock in northern Idaho.

During maximum glaciation, the ice was thick enough to pass over the highest peaks of the Selkirk and Cabinet Ranges at elevations of more than 6,000 feet. This required an ice sheet to be more than 4,500 feet thick in the vicinity of Sandpoint. The ice may have been more than 2,000 feet thick at the southern end of Lake Pend Oreille during maximum glaciation.

Alpine Glaciation in Idaho

From 7,000 to 25,000 years ago, alpine glaciation was widespread in the higher elevations of the state. At least two periods of major glaciation are evident in Idaho. The last stage of alpine glaciation occurred about 4,000 years ago. This glacial action was relatively minor as glaciers existed only in the highest mountains of the State. During interglacials, meltwater rivers deposited vast amounts of sand and gravel in the mountain valleys, at the margins of ranges, and the Boise valley. Evidence of Pleistocene glaciation can be seen in mountainous areas at elevations as low as 5,000 feet above sea level in Idaho.

Glacial Lakes and Floods

Large quantities of glacial meltwater had a dramatic effect on the landscape. Much rock debris was transported by water and deposited in valleys. Many floods were caused by glacial ice impounding water and then bursting. Huge catastrophic floods were caused in such a manner and drastically eroded the landscape.

During and immediately following the ice age, the streams of Idaho carried much more water than they do now. Larger streams and rivers could transport a much greater sediment load, mostly of glacial debris. At this time, the abundance of water caused large lakes to form in closed basins. One of the largest of these lakes was ancient Lake Bonneville, once covering more than 20,000 square miles with a maximum depth of more than 1,000 feet. The Great Salt Lake is a remnant of this lake.

Ancient glacial Lake Missoula was created by an ice lobe forming a dam near the Idaho-Montana border. By melting or erosion this dam was suddenly removed and great floods were released throughout the northwest. Glacial debris left by the retreat of the great glaciers dammed streams and formed many modern lakes in Northern Idaho, including Hayden, Spirit and Twin Lakes. Pend Oreille Lake was formed in a similar fashion by glaciers eroding the lake basin and glacial debris damming the south end.

Loess of Eastern Idaho

Loess is homogeneous, unstratified, predominately silt-sized material transported to the site of deposition by wind. It is very porous and can maintain a very steep slope. Approximately 30 percent of the United States is mantled with loess including much of the Snake River Plain. In fact loess represents most of the soil found on Pleistocene basalt flows in the Snake River Plain. Loess generally mantles the landforms with a deposit of equal thickness over ridges and valley bottoms much like a snow fall. However, later erosion by both wind and water tend to transport the loess to topographically low areas. Field evidence will always show that loess is not derived from the underlying soil or rock.

Russell (1902) first mentioned the fine, yellowish white silts and their similarity with the loesses of China and the Mississippi Valley. He noted that these windblown silts were not derived from the underlying lava flows but were transported in by wind. Almost all the soils of southeast Idaho were at least modified by additions of loess, even though primarily derived from the other sources. Loess soils are characterized by uniform grain size and a good capacity for moisture holding.

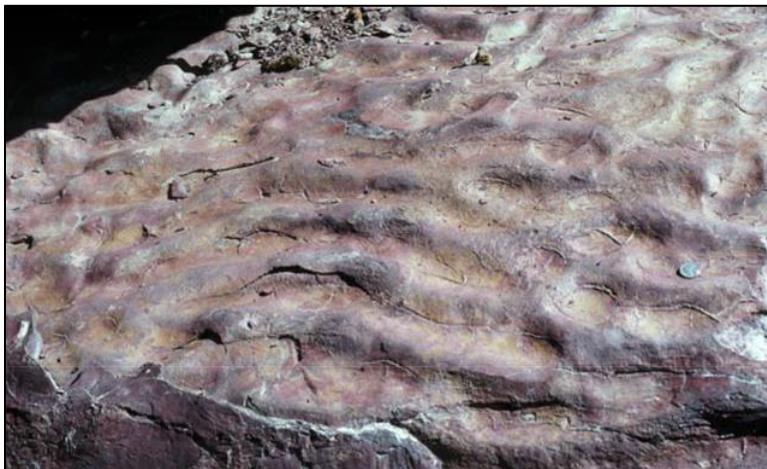
Most of the basalt of the Snake River Plain has a loessal cover more than 3-feet thick. However, Holocene basalt has an almost complete lack of loess. The Pleistocene-age deposits on the older basalts have a variable thickness. The swales may have thickness of up to 100-feet thick while the pressure ridges and plateaus may have loess less than a meter thick.

When loess deposits reside on basalt flows, which are topographically high relative to surrounding areas, it is good evidence that all the loess is windblown silt rather than including or consisting of decomposed rock below or contaminated by fluvial sediments. The effect of loess deposition on land forms, particularly the rough hummocky surface of basalt flows is to smooth the surface by filling in the low areas but also mantling the high areas. During Holocene time (last 10,000 years) there was very little deposition.

Geologic Provinces, Terranes, Structures and Events

Belt Supergroup

The Belt Supergroup sediments were deposited in an intracratonic basin between 1450 and 1250 million years ago. However, there is some possibility that the Belt Supergroup could have been deposited as recently as 900 million years ago. These sedimentary rocks crop out over vast areas of northern Idaho, western Montana, eastern Washington, and western Canada. They reach thicknesses of 50,000 feet near the Washington-Idaho line.



Ripple Marks in Precambrian Belt Supergroup sedimentary rocks. Ripple marks are suggestive of a shallow water origin. Glacier National Park.

The sand grains deposited in the basin originated in a land mass that rifted away. The portion that rifted to the southwest away from the North American craton is possibly Australia. To accommodate the thick sequence of sedimentary strata, the Belt basin may have subsided as much as 10 miles over a 200-million-year period.

Much of the Belt Supergroup was tectonically moved eastward along thrust faults during Mesozoic time. The western portion of these rocks was intruded and metamorphosed by the Idaho Batholith.

Most of the Belt sedimentation appears to be of continental or lacustrine (lake) origin rather than marine. The Prichard Formation of the lower Belt consists of turbidites and deltaic sediments. The Ravalli and Missoula Groups consist mostly of fluvial strata which were deposited on broad, unchanneled sand and mud flats. The middle Belt carbonate consists of fine-grained carbonate and terrigenous clastic mud deposited in alternating siliclastic (sandy) and carbonate cycles.

Major Stratigraphic Divisions

Belt rocks are separated into four major stratigraphic Divisions: the Missoula Group with 30,000 feet of strata; the Helena-Wallace Formations with 10,000 feet of strata; the Ravalli Group with

6,600 feet of strata; and the Prichard Formation with 23,000 feet of strata.

Shallow Water Depositional Environments

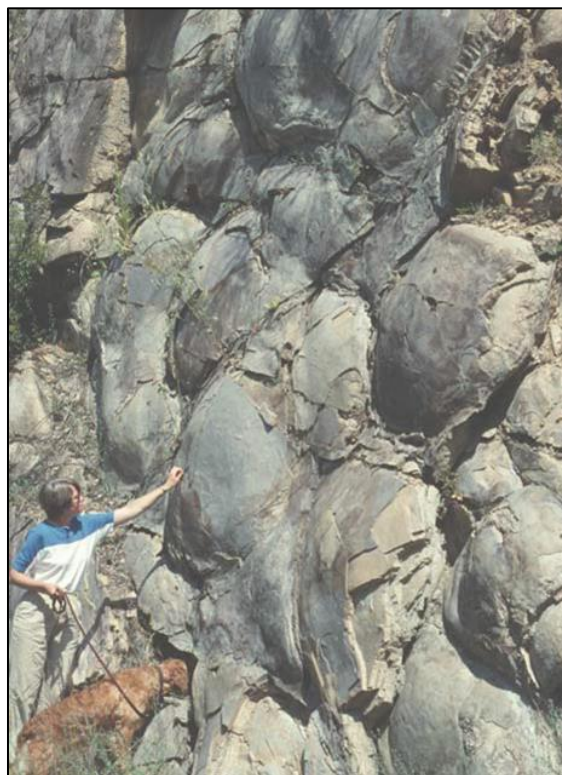
Sediments collected in the Belt basin exhibit substantial evidence of shallow-water environment of the Belt Supergroup. These include fossil stromatolitic algae, mud cracks, small-scale cross-bedding, ripple marks, salt crystal casts, flute casts, groove casts, load casts, and mud-chip breccia.

Stromatolites

Fossil forms of ancient algae or former marine plant life are called stromatolites. These stromatolites are similar to the modern blue-green algae; and in the outcrop they consist of alternating bands of light and dark-colored mineral matter arranged in swirling patterns. Fossilized algae have nearly spherical or ellipsoidal structures ranging from the size of a football to a large reef.

All the algal forms are collectively called stromatolites or, when forming massive rock, they are called stromatoliths. The genus name for many of the forms in this area is *Collenia*. *Collenia* lived in protected intertidal flats of the ancient Precambrian Beltian sea.

The algae probably grew many miles from shore but in water ranging from the tidal zone to 100 feet deep. The water had to be shallow enough so as to allow sunlight to reach the algae. Blue green algae, like the stromatolites, took carbon dioxide from sea water and sunlight and released oxygen as a waste product through the process of photosynthesis. This process was not only a major factor in producing an oxygen-rich atmosphere which made it possible for oxygen-consuming life forms to exist on Earth, but also caused large quantities of calcium carbonate to be deposited on the sea bed. When carbon dioxide is removed from the seawater by algae, it generates a chemical reaction that forms fine particles of calcium carbonate.



Large, well developed domal stromatolites exposed along steeply dipping bedding surface of Belt argillite.

Accreted Terrane in Western Idaho

Most of the pre-Cretaceous rocks west of the Idaho Batholith in west-central Idaho and east-central Oregon are oceanic or island arc assemblages. These rocks were formed offshore in island arcs and adjacent basins and were accreted to the North American continent between Late-Triassic and mid-Cretaceous time. This means that before Jurassic time, the west coast of North America was situated near the Idaho-Oregon border.

The Salmon River Suture Zone

The Salmon River Suture zone forms the Paleozoic and Mesozoic border of the North American continent for approximately 300 miles extending from the South Mountain area of southwestern Idaho to approximately Orofino where the boundary abruptly trends west. At this border, oceanic island arc rocks to the west in what is now Oregon and Washington were rafted against the old North American continental rocks of central Idaho. The rocks to the east of this suture area are Belt Supergroup and pre-Belt rocks of Proterozoic age. These two settings are separated by the strontium-isotope line. All the plutonic rocks west of the dashed line have low initial ratios (<0.704); whereas, all rocks to the east of the line have high initial ratios (>0.706). This change in ratios occurs in less than a distance of 6 miles. The strontium-isotope line therefore represents the suture line where the accreted island arc assemblages were welded to western North America.



View looking north-northwest across Little Salmon River in west central Idaho. Lower half of canyon consists of accreted terrane; upper half consists of Columbia River Basalt flows. The layering of the basalt flows forms a syncline and the contact between the basalt and the underlying accreted terrane is an unconformity. W. Hamilton, 1963.



Volcanoclastic greenstone of the accreted terrane, west central Idaho.

Accreted Terranes

The Blue Mountain province of northeastern Oregon and western Idaho on the eastern side of the Salmon River Suture zone consists of 4 east-west-trending terranes. These terranes from north to south include (1) the Wallowa Terrane (a volcanic arc active from Late Paleozoic to Late Triassic time), (2) Baker Terrane (partly *mélange* or deformed material in the subduction zone as the terrane was accreted eastward beneath the Olds Ferry volcanic arc), (3) the Izee Terrane (intra-arc and fore-arc basin deposits), and (4) the Olds Ferry Terrane (a west-facing, intra-ocean volcanic island arc active from the Middle Triassic to Middle Jurassic time).

Deformation and Time of Accretion

The accreted terrane was deformed in the Late Triassic and again in the Late Jurassic. The Late Triassic deformation occurred following deposition of most of the volcanic rock units. The time of the accretion is estimated to have occurred between 95 and 120 million years ago.

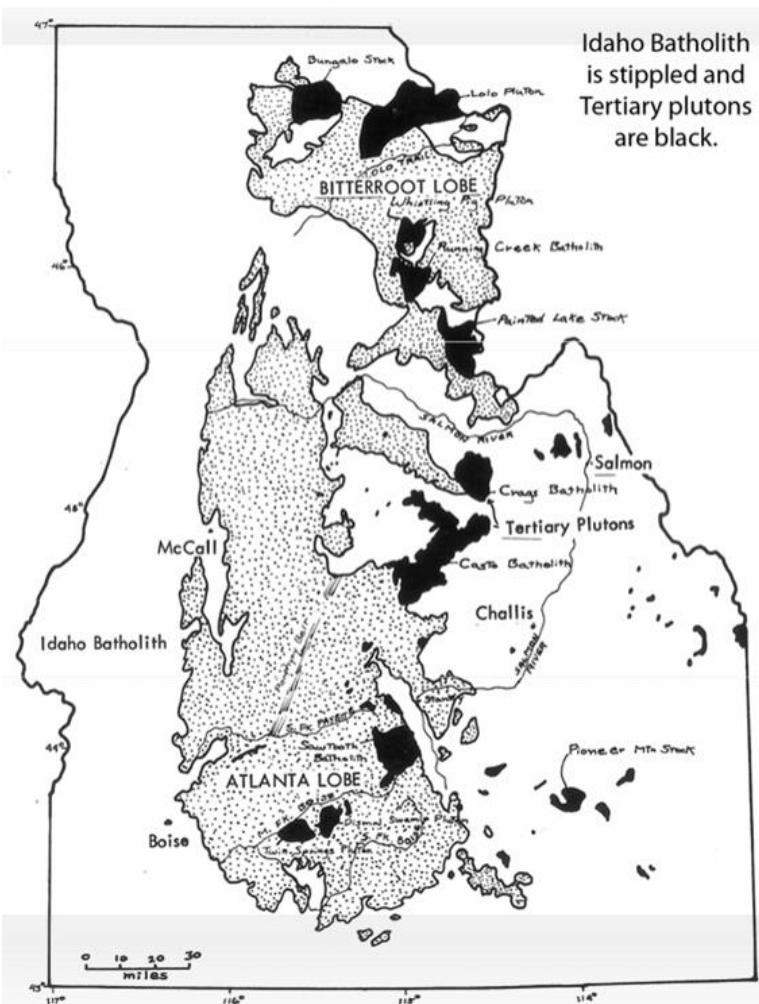
Deformation and metamorphism of the Riggins Group at the contact with continental rocks occurred at that time. However, the accretion process probably occurred over a period of time ranging between Late Triassic and mid-Cretaceous. During this time and for a period afterwards, the Idaho Batholith was formed by magmas generated from subduction of the eastward-moving plate.

Idaho Batholith

The Idaho Batholith is a composite mass of granitic plutons covering approximately 15,400 square miles in central Idaho. The outer perimeter of the batholith is irregular and 200 miles long in the north-south direction and in plain view it has an hour glass shape. It is approximately averages about 75 miles wide in an east-west direction.

Age of Batholith

Armstrong and others (1977) called the northern part of the hourglass the "Bitterroot" lobe and the southern part the "Atlanta" lobe. He also proposed that most of the



southern lobe was emplaced 75 to 100 million years ago (Late Cretaceous); the northern lobe was emplaced 70 to 80 million years ago. Armstrong (1977) further noted that older plutons of Jurassic age occur on the northwest side of the Bitterroot lobe and many Eocene plutons have intruded the eastern side of the Atlanta lobe of the batholith. On the western side of the batholith, there are more mafic plutons (quartz diorites or tonalites) than to the east.

Radiometric dates and field relationships, where plutons of the batholith cut older rocks, restrict the age of the Idaho Batholith to an interval between 180 million years ago (Late Triassic) to 45 million years ago (Eocene); however, the dominant interval of emplacement was Early to Middle Cretaceous. There is a general west-to-east decrease in age for plutons of the batholith.

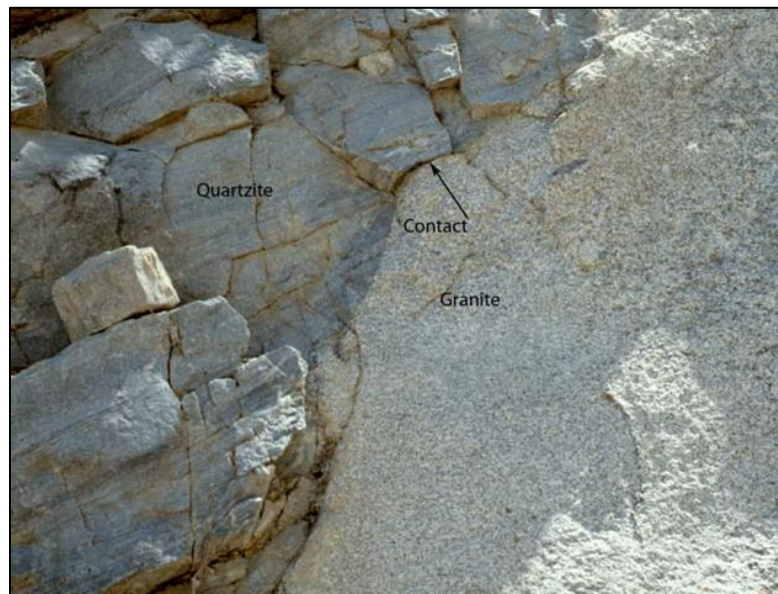
Bitterroot Lobe

Hornblende-biotite tonalite and quartz diorite plutons were emplaced at mesozonal levels in the western and northwestern margin of the Bitterroot lobe. Later, during Paleocene time, plutons of foliated granodiorite were intruded. These plutons of sedimentary-type granites are chemically similar and represent most of the bitterroot million years ago.

Dike Rocks in the Batholith

Pegmatite and aplite dikes were formed during the late stages of each plutonic intrusion. The mineralogy of the dikes is similar to the enclosing intrusive. In some cases, pegmatite dikes cut aplite dikes and in other cases the reverse is true. These dikes appear to be concentrated near Eocene plutons and occur in the northeast-trending, trans-Challis fracture zone.

The Bitterroot lobe of the Idaho Batholith contains numerous mafic dikes which make up about 20 percent of the total rock. These tabular dikes average about 8 feet thick and trend east-northeast.



Granitic batholith on right intruded into quartzite. The contact is discordant (cuts across the layering in the quartzite).

Field Identification of Granitic Rocks

Granitic outcrops of the Idaho Batholith are easily recognizable in the field.

Under close inspection, granite has a salt and pepper appearance with the dark minerals of biotite

mica and hornblende and light minerals of plagioclase and quartz. The constituent minerals are up to an inch or more in diameter and can readily be identified without a hand lens. Of course, many minor accessory minerals are too small to be identified with the unaided eye.

Weathered outcrops of granite have a distinctive appearance and can in some cases be identified at a mile or more distance. Coloration tends to be very light gray to very light tan, and in some places chalk white due to leaching by hot water. Outcrops are generally smooth and rounded due in part to surface weathering by granular disintegration and in some cases exfoliation where layer after layer is removed from the surface. Most exposures are cut by one or more sets of fractures which may give the outcrop a blocky appearance. Granite and basalt are among the easiest rocks in Idaho to identify.

Tertiary Plutons of the Batholith

Approximately 20 percent of the Idaho Batholith is composed of granitic rocks younger than the Mesozoic. The largest Tertiary-age plutons are batholith in size and are crudely aligned along a north-trending belt in the east-central part of the batholith. These plutons have long been known to be Tertiary age, if not Eocene, because they cut or intrude older rocks including the Idaho Batholith. Some of the plutons even cut Challis volcanic rocks.

More than 40 plutons of Tertiary age (42 to 46 million years old) have been identified in or near the Idaho Batholith. These intrusions range in size from dikes or small stocks to batholiths. The small stocks and dikes may be apophyses (narrow upper portion) of larger batholith-sized plutons at depth.

Pluton Characteristics

The Tertiary granites have a pink color caused by pink potassium feldspar, whereas rocks of the Idaho batholith tend to be gray. The composition range is fairly narrow ranging from quartz monzonite to granite. Mirolitic (gas) cavities are characteristic of the Tertiary plutons. These gas cavities indicate a shallow level (epizonal) of emplacement. The Tertiary plutons may have been emplaced within 3 to 4 miles of the surface, whereas the Idaho batholith may have been emplaced more than 6 miles from the surface. Euhedral crystals (well-formed crystal faces) of smoky quartz and feldspar commonly occur in the mirolitic cavities.

The Tertiary granitic rocks contain twice as much uranium and thorium than is found in the Mesozoic batholith. Because of this difference, a gamma ray spectrometer can be used to map the contacts between the Tertiary plutons and the surrounding batholith.

Some of the large plutons such as Bighorn Craigs, Sawtooth and Casto Batholiths have a well-developed vertical jointing. This jointing causes a very distinctive and easily recognizable ragged

topography as is exemplified by the Sawtooth Mountains near the town of Stanley.

The potassium feldspar crystals are not as large in the Tertiary plutons as they tend to be in the batholith; this is probably caused by the shallow level of emplacement and quick cooling in the Tertiary rocks.

Challis Volcanism and Tertiary Plutons

The Challis volcanic field of east-central Idaho is very likely part of the same magmatic event that caused the Tertiary granitic plutons. The Challis volcanics represent the portion of the magma derived from the epizonal plutons that managed to break through the surface and be extruded as flows and tuffs.

Smoky Quartz Crystals

Bennett (1980) observed that smoky quartz crystals in miarolitic cavities are characteristic of the Tertiary granitic plutons, with perhaps the best specimens coming from the Sawtooth Batholith. Tertiary granites have twice the background radiation as the Cretaceous Idaho Batholith. This makes it possible to develop smoky quartz crystals which must be formed in a radioactive environment. These cavities tend to be preferentially confined to zones which may be caused by tension joints or fractures. During cooling of the granitic mass, these joints are caused by contraction.

Most smoky quartz crystals are less than one inch in length. Smoky quartz crystals left on the surface for several years will turn clear because of exposure to sunlight. Many miarolitic cavities also contain euhedral crystals of feldspar and aquamarine (beryl).

Challis Volcanism

The Challis volcanic rocks are a thick series of volcanic flows and tuffs that cover a large part of east-central Idaho. Interbedded with rhyolitic volcanic flows and tuffs are lake bed sediments as well as fossiliferous sediments formed from other processes. Fossil plant species as well as radiometric age dating indicate an Eocene age. The volcanism started about 51 million years ago from a variety of widely-separated vents and continued until about 40 million years ago.



Colorful Challis volcanic rocks in east-central Idaho.

Extent of Volcanism

Challis volcanics now cover approximately 1,900 square miles of east-central and south-central Idaho. At the end of the eruptive period, the Challis volcanic rocks were so widespread that they covered more than half of Idaho. Thickness of the Challis volcanics is variable and in several places is known to be more than 10,000 feet -

Rock Types

There were many large-volume eruptions of intermediate lava from numerous centers. Both basaltic and rhyolitic lavas were erupted. Lavas and volcanoclastic rocks of intermediate to mafic composition were extruded from widely-scattered vents. These eruptions were nonexplosive and of small volume. Rhyolite lavas and domes and rhyolitic ash-flow tuffs were erupted from caldera complexes.

Calderas

The numerous ash-flow tuffs are generally believed to be derived from calderas. An example of one of these calderas is the Twin Peaks Caldera. It is a roughly-circular collapse structure, 12.5 miles in diameter and established about 45 million years ago. The calderas are associated with large geothermal systems that operated throughout the Challis event. The heat in the crust from a pluton below would have caused deep circulation of water and provided an ideal environment for ore deposits to form. Because of this, calderas have important significance to exploration geologists.

Exposures Near Challis

The base upon which the Challis volcanics rest is a rugged mountainous surface of Precambrian and Paleozoic sedimentary rocks. While driving along highway 75 in the vicinity of Challis, it is possible to see many excellent exposures of the Challis volcanics as well as the irregular base on which they lie. The air-fall tuffs in this area have a soft appearance, are easily eroded and are variable in color, including lavender, light green and several other pastel colors.

Owyhee Mountains and Plateau

Owyhee Mountains

The northwest-trending Owyhee Mountains have a steep northeast side adjacent to the western Snake River Plain. This steep side was formed by normal faulting with some downwarping as the western plain subsided. Granite in the Owyhee Mountains is related to the Idaho Batholith in terms of origin, age and composition. This granite contains Paleozoic or earlier roof pendants of metasedimentary rocks. South Mountain consists of a pendant of older metamorphosed rocks. Challis volcanics of Eocene age crop out in the Poison Creek area and Oligocene volcanic rocks are found in the Salmon Creek area.

The highest Owyhee Mountains exceed 8,000 feet and during the last glacial period these mountains were extensively glaciated. Glacial cirques, U-shaped valleys, and moraine deposits are common.

The Owyhee Plateau

The Bruneau-Jarbridge low plateau is a rhyolite eruptive center that was active about 12.5 to 8 million years ago. When it settled into a broad basin, huge volumes of rhyolite lava filled the basin. Later, fluid basalt was erupted in the basin and significantly smoothed the topography.



Deeply incised Owyhee River on the Owyhee Plateau, southwest Idaho.

The Owyhee-Humboldt plateau was the site of an earlier eruptive center which was active about 14 million years ago. This basin was also covered and smoothed with flows of basalt. The canyons of the Owyhee, Bruneau, and Jarbridge Rivers cut about 1,000 feet through many layers of volcanic rocks.

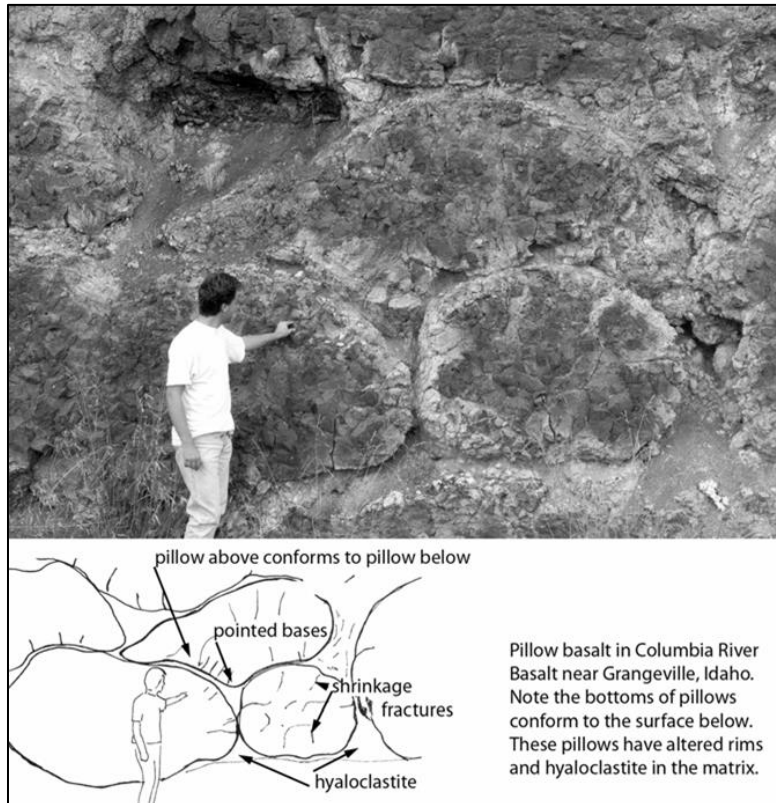
Bruneau-Jarbridge Eruptive Center

Volcanic flows, exposed by the canyon of the Bruneau River, were erupted from the Bruneau-jarbridge eruptive center. This eruptive center is about 59 miles long and lies southeast of the scenic Bruneau Canyon. Numerous ash flows, lava flows and basalt flows were erupted during Miocene and Pliocene time. The Banbury Basalt was erupted from shield volcanoes and filled in the low areas. The Cougar Point Tuff consists of densely-welded, ash-flow tuffs and younger rhyolite flows about 10 to 12 million years old. Altogether there are nine or more of these welded tuff cooling units which filled a large basin. The eruptive center was named by Bonnicksen after the Bruneau and Jarbridge Rivers which have cut canyons in the area. The canyons expose both silicic and basalt flows.

Columbia River Basalt Group

The Columbia River Basalt Group covers an area of approximately 77,000 square miles in Oregon, Washington and western Idaho. It forms the largest Cenozoic basalt field in North America and has an estimated total volume of approximately 90,000 cubic miles. Radiometric age determinations indicate that the group was extruded during a period from 17 to 6 million years ago. About 99 percent of the volume was erupted during a 3.5 million year period from 17 to 13.5 million years ago.

This episode, which may be linked to the Yellowstone hotspot covered most of the States of Oregon, Washington and Idaho. More than 175 separate flows of the Columbia River Basalt Group total more than 3 miles thick in vicinity of Yakima, and have a volume of more than 42,000 cubic miles. Individual flows of more than 500 cubic miles cover a large part of the States of Oregon, Washington, and Idaho with some extending more than 470 miles from their source. The Group was extruded during a period from 17 to 6 million years ago from northwest- trending fissure dikes; however, most of the total volume was extruded between 13 and 17 million years ago. At least 7 of the stratigraphic units are separated by erosional unconformities and many of them have saprolite or paleosols (fossil soils) indicating a significant time period between flows.



Pillow basalt in Columbia River Basalt near Grangeville, Idaho. Note the bottoms of pillows conform to the surface below. These pillows have altered rims and hyaloclastite in the matrix.

The main phase of the Columbia River Basalt eruption began in southeastern Oregon from 16.6 to 15.0 million years ago—or about 1.5 million years. This represents the greatest eruptive volume of continental basalt on Earth during the past 30 million years. These flows include the Steens, Innaha, Grande Ronde, and Picture Gorge Basalt.

Source Fissures and Dikes

Before the western Snake River Plain rift basin formed, widespread and very voluminous volcanism was occurring in the region. Columbia River Basalt flows were erupted from north-northwest-trending fissures in northeastern Oregon, eastern Washington, and western Idaho. An average flow of Columbia River Basalt was approximately 100 feet thick, although some were more than 200 feet thick.

Three Embayments in Idaho

The first flows of Columbia River Basalt were erupted over a landscape of rolling hills. Each successive flow filled in the low areas until a flat-surfaced plateau was formed. Three lobes of

Columbia River Basalt extended eastward into Idaho from the Columbia Plateau covering topographically low areas: the Weiser embayment, the Clearwater embayment, and the St. Maries embayment. Individual flows and sequences of flows can be traced for many miles. They can be recognized on the basis of a variety of chemical, petrological, stratigraphic, and magnetic polarity criteria. The average thickness of the flows ranges from 50 to 100 feet but some are more than 400 feet thick in basins.

The eastern plateau has been uplifted almost 6,600 feet allowing deep erosion by the Snake, Salmon, Clearwater and other rivers. Consequently the entire section is revealed in several localities of west-central Idaho.

According to Wood and Clemens (2002), there is a thick layer of basalt lava situated beneath the sediments in the western Snake River Plain. They observed that the deepest drill hole in the plain near Meridian penetrated 0.7 km of lacustrine sediment and a thick section of basalt and tuffaceous material. This basalt is geochemically within the range of Columbia River basalts, particularly those that erupted 14 million years ago in eastern Oregon and is similar to the basalt in the Weiser embayment. On the basis of borehole data, they estimated that 6500 feet of interbedded basalts and tuffaceous sediments underlie the Idaho Group sediments.

It seems likely that the isostatic adjustment caused by the emplacement of such a thick sequence of basalt would have contributed to the depression of the western Snake River Plain. For the same reason, the adjacent land (just beyond) around the basalt flows would have experienced uplift.

Snake River Plain

The Snake River Plain is a prominent depression across southern Idaho extending 400 miles in an east-west direction. It is arc shaped with the concave side to the north. The width ranges from 50 to 125 miles with the widest part in the east. This physiographic province was originally referred to as the Snake River Valley or the Snake River Basin. However, in 1902 T. C. Russell re-designated the province the Snake River Plains. Later the name was changed to Snake River Plain to convey a sense of uniformity throughout the province.

The subsurface of the plain is known because of thousands of water wells and several deep exploration wells for geothermal resources and oil and gas. Geophysical surveys have also yielded much information on the subsurface.

The Snake River Plain originated and evolved during the last 17 million years. Although at first glance, the Plain appears to be a homogenous feature, the western plain is quite different from the eastern plain in terms of structure, lithology and age. The western Plain trends northwest-

southeast and the eastern plain trends southwest-northeast. Generally, the relief of both the western and eastern Plain is low with some relief provided by a scattered cinder mounds and shield volcanoes. The western Snake River Plain is a structural basin formed by large gravity faults at both the northeast and southwest boundaries; the eastern Plain is a downwarp with no solid evidence of boundary faults.

The Snake River

Most of the streams in southern Idaho flow into the Snake River and then down the Columbia River and into the Pacific ocean. To the south of eastern Idaho is the Great Basin, an area of internal drainage covering western Utah, extreme southern Idaho and eastern Nevada. The ultimate base level for the Great Basin is the Great Salt Lake. No stream from the Great Basin reaches the ocean.

The Snake River starts its journey across the plain in the vicinity of Jackson Hole, Wyoming and follows the southern edge of the Quaternary basalt flows. The reason for this is the basalt is topographically higher than the adjacent Snake River. This phenomenon is characteristic for streams and rivers in the Snake River Plain. Many of them, such as the Big Wood River, were moved by the spread of lava flows and tend to conform to the margins of the flows.

The Snake River did not always have its headwaters in Wyoming. As the North American plate slowly moved southwesterly over the fixed hot spot, the topography over the hot spot uplifted to a high plateau. When the movement of the North American plate placed the fixed hot spot under what is now the central Snake River Plain about 10 million years ago, the central Snake River Plain was a high volcanic plateau. This high plateau moved progressively to the northeast across the eastern Snake River Plain where it is now situated at Yellowstone. As the surface land on what is now the Snake River Plain subsided from cooling behind the northeast-moving plateau, the drainage systems in the Snake River Plain were rerouted. Instead of streams flowing radially off of the advancing plateau, the drainage assumed its present westward flow (Link and Phoenix, 1994).

The term "Lake Idaho" or "Idaho Lake" is applied to the lake responsible for the lacustrine sediments that are Pliocene or more recent in the western Snake river Plain. Lake Idaho was probably not a continuous lake with a constant shoreline but was modified greatly through the last 10 million years from volcanic action and available glacial melt water.

The biogeographic evidence, which is based on divergent speciation of Pliocene molluscan faunas in southern Idaho and Utah, indicates that a drainage divide existed between the western and eastern Snake River Plain. Before the Quaternary, streams from southeastern Idaho drained into the Bonneville basin. Only during the Quaternary did the Snake River and its tributaries in southeastern Idaho become connected to southwestern Idaho and the Columbia to form a single

drainage basin.

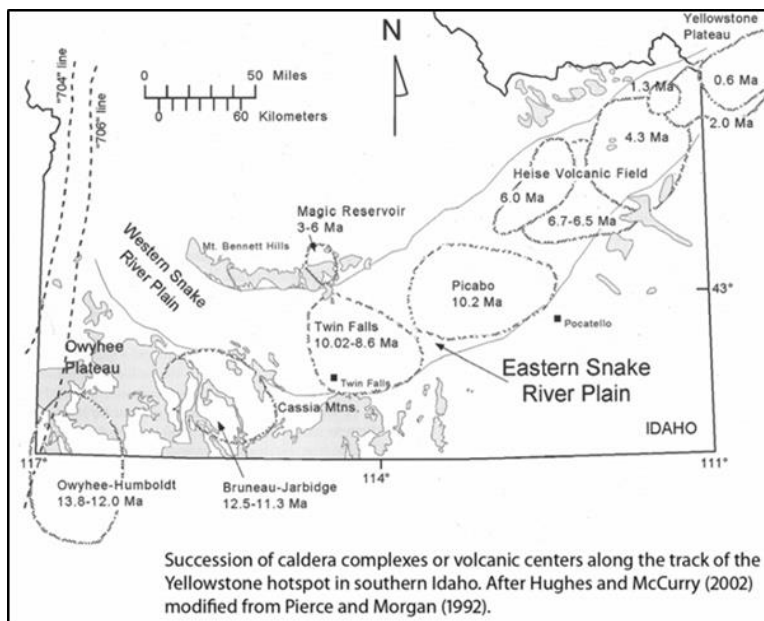
Capture of the Western Snake River Plain Drainage

In 1898, Lindgren first recognized that sediment-filling of the western Snake River Plain basin was followed by river entrenchment. Wheeler and Cook (1954) proposed that the drainage of the western Snake River Plain was captured near the Oxbow of the Snake River in Hells Canyon. They suggested that the Snake River eroded south and captured Lake Idaho. The Draining of Lake Idaho put so much water through Hells Canyon that it eroded a channel deeper than the Salmon River. As a result, the Salmon River became a tributary of the Snake.

Track of the Yellowstone Hotspot

The Yellowstone hot spot left a 430-mile-long, northeastward-trending track consisting of time-transgressive caldera-forming volcanism. This silicic volcanism appears to have started about 17 million years ago in a north-central Nevada, and then progressively moved in an east-northeast direction across northeast Nevada, southeast Oregon and southern Idaho. The time-transgressive silicic volcanism is explained by the west-southwest movement of a

60-mile-thick lithosphere (the North American plate) over a stationary thermal mantle plume or hotspot at the rate of 1 inch per year. The rhyolitic rocks are progressively younger towards the east starting with dates of 11.3 million years old at the Bruneau Jarbidge eruptive center to about 4.3 million years old at the east end of the Snake River Plain. The most recent rocks on the topographically high Yellowstone Plateau were erupted as recently as 0.6 million years ago. The Basin and Range Province, the Columbia River Basalt, and the western Snake River Plain all originated about 17 million years ago and may be linked to the Yellowstone hot spot.



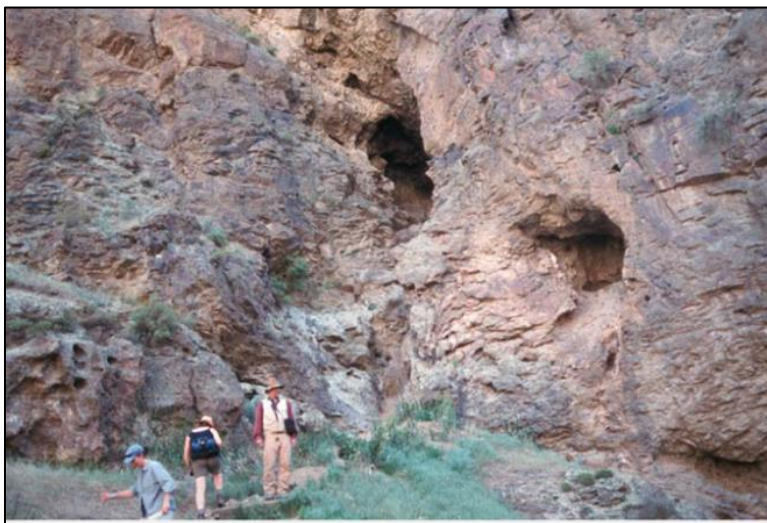
Western Snake River Plain

The western Snake River Plain is 30 to 43 miles wide and trends northwest. It is a fault-bounded basin with both the land surface and the rock layers dipping towards the axis of the plain.

The basin is filled by interbedded volcanic rocks and lakebed sediments of Tertiary and Quaternary age.

Northwest-Southeast-Trending Faults.

Lindgren (1898b and Russell (1902) first proposed that high-angle faults caused the escarpment along the northern side of the western Snake River Plain. Later, Malde (1959) came up with compelling evidence for a boundary faults on the northeast side of the western plain. Subsequently, a number of investigators have published evidence of normal faults displaced down toward the center



Large lithohyal (air) cavities in the interior of the Jump Creek rhyolite lava flow near Jump Creek Falls, southwestern Idaho.

of the plain on both the northeast side and the southwest side of the western Snake River Plain. The vertical displacement along the northern faults is up to 9,000 feet with deformation beginning in the Miocene and continuing into the Pleistocene.

The deep wells drilled in the western plain show interbedded basalt and sediment. One well located about 43 miles south of Boise passed 11,150 feet of alternating sediment and volcanic rocks before it terminated in granite. This granite may be a southern extension of the Idaho batholith.

Geophysical surveys used to interpret the Snake River Plain include gravity surveys, magnetic surveys, seismic refraction profiles, thermal gradient measurements, and heat-flow measurements. There is a gravity high over the plain. This gravity high coincides well with the topographic low. The gravity high anomaly over the western plain is interpreted as indicating a thin upper crust. Seismic refraction data also support this interpretation. Seismic refraction profiles indicate that the total crust under the plain is more than 25 miles thick; however, the upper crust is thin under the axis of the plain. Heat-flow measurements indicate a high heat flow anomaly along the margins of the plain and a relatively low heat flow in the central part of the plain.

Huge Rhyolite Lava Flows and Pyroclastic Deposits

Most of the rocks southwest of the western Snake River Plain consist of rhyolite lava flows and pyroclastic deposits. There are rhyolite lava flows along the Owyhee front and along the Boise

front. In fact, these flows along the Owyhee front are scientifically very significant because they represent some of the largest on Earth. Many pyroclastic deposits have a typical depositional sequence. For example, at the base may be a coarse air-fall tuff, followed by a pyroclastic surge deposit, then a pyroclastic flow, and capped by an ash-fall deposit.

All volcanoclastic material ejected from a volcano is called tephra and is represented by fragments of volcanic glass or pumice, glass shards, crystals (most commonly quartz and feldspar), and fragments of lava from earlier eruptions. Tephra includes dust, ash-sized material, as well as larger fragments.

Vesiculation and Explosive Volcanism

Pyroclastic (fragments of rock formed by volcanic eruption) deposits are formed by explosion out of a central vent. Vesiculation is the separation of volatile phases in magma. This causes a great increase in volume and may lead to explosive volcanism with the associated pyroclastic debris. Low-silica magmas such as basalt contain less water so they are less susceptible to explosive activity.



Montini Volcano, dated at 1.64 Ma, has an original Maar from the late-stage magmatic components (Godchaux and Bonnicksen, 2002). It has been dissected by both the Snake River and Sinker Creek which has provided exposures of the ring dike segments and the crater-filling tuff and lava. Photo by H.E. Malde, 1978; U.S. Geological Survey Photo Library.

- *Calderas*. On a significantly larger scale than craters, calderas also represent the surface expression of a collapsed volcano. Craters typically are less than 0.5 miles in diameter; whereas, calderas are commonly more than 6 miles in diameter. Calderas generally form in response to the withdrawal of large quantities of magma from an underlying magma chamber and the unsupported layer of rock over the chamber collapses. Calderas that generate ash-flow tuff eruptions collapse along arcuate ring fractures. After a sufficient volume of magma is expelled, the unsupported roof subsides into the chamber. The Bruneau-Jarbidge eruptive center is a complex of calderas and the source of huge volumes of ash-flow tuffs.
- *Obsidian*. Obsidian forms when magma of rhyolitic composition cools so fast that crystallization of minerals is not possible. Typically, volcanic glass is a lustrous, glassy black or reddish-black rock with a conchoidal fracture giving it very sharp edges.

- *Jasper and Other Secondary Deposits.* Secondary silica, in the form of opal, chalcedony, or jasper, were deposited in fractures, gas cavities, spherulite shrinkage cavities, between breccia fragments, and other pore spaces in the rhyolite flows. These silica deposits formed after the lava was formed, but before it cooled, by hot water leaching silica from the rhyolite mass and precipitating it in the higher portions of the flow where the rock was cooler. The Bruneau jasper deposit is a high quality, commercial-grade picture jasper of red, brown, and tan colors. Most of this jasper occurs as jasper-filled spherulite shrinkage cavities and as fracture filling.

Eruptive Centers

The Owyhee-Humboldt, Bruneau-Jarbridge, and Twin Falls eruptive centers were successively formed between 14 and 8 million years ago. Each eruptive center is a large structural basin filled primarily with volcanic rocks and minor younger sedimentary rocks. In a few places faults are exposed that indicate downward displacement (subsidence) of the basin interiors. The eruptive centers are like calderas but much larger; they were probably formed by several overlapping calderas—all aligned along the axis of the Snake River Plain.

Rhyolitic volcanism in the western Snake River Plain began along the southwestern margin of the rift starting at about 11.7 Ma. Between 12 and 10 Ma, there was an enormous volume of high-temperature ignimbrites and large rhyolite lava flows erupted in southwestern Idaho. The Bruneau-Jarbridge area experienced the most voluminous eruptions in Earth history. Buried under the Bruneau-Jarbridge eruptive center is a caldera complex. When these calderas collapsed, the huge rhyolitic eruptions occurred.

Old Basalt 9.5 to 7 Ma

During two periods, basalt erupted 9-7 million years ago and 2.2-0.4 million years ago in Lake Idaho in the western Snake River Plain. Depending on the presence of water, eruptions ranged from effusive to very explosive. Eruptions yielded subaerial (surface) flows, pillow deltas, water-affected basalt, fragmental basalt deposits and basaltic tuffs. Most of the basalt was erupted during the 9-7 million-year period and, for the most part, was erupted in deep water. The most recent volcanism in the western Snake River Plain has produced small shield volcanoes, subaerial flows, and cinder cones.

The changing lake level through time resulted in a great variety of basalt eruptions and products. Shoreline features of this paleolake are still preserved at approximately 3800 feet for the high-stand elevations.

Summary of Historical Views on Lake Idaho

The first geologists to explore and report on the geology of the central and western Snake River Plain recognized the sediments represented a large Tertiary lake. Cope noted the similarities of

the fossil fish in the Lakebed sediments to those found in Lake Bonneville and Lake Lahontan. He named the ancient water body Lake Idaho. Later geologic studies by Russell (1902) and Lindgren and Drake (1904) further established that a large lake once existed in the western Plain.

General Characteristics of Lake Deposits

Lakes generally form in a landlocked basin or impoundments that eventually fill up, but may subside as new sediments are added. Depending on the size of the lake, deposits may range from less than a meter to thousands of meters thick. Lakes form a series of deposits that are concentrically zoned from coarse sandstones at the outer margins and grading to mudstones at the center. Also, because lakes gradually fill over time, the sequence will also show an upwards coarsening of grain size from laminated shales at the bottom to cross-bedded sandstones at the top. Evidence of a lacustrine environment of deposition include freshwater fossils, especially plant material in fine-grained laminated sediments, oscillation ripples instead of current ripples, and a lack of tidal structures.

Evidence of Early Large Lake

Godchaux and Bonnicksen (2002) point out that when the early rhyolite eruption occurred along the southwest side of the plain between 11.7 and 11.0 Ma, a lake already existed. They observed that the distal parts of many of the rhyolitic lava flows had interacted with large amounts of surface water, presumably, the earliest time of Lake Idaho. They also noted that most of the early rhyolitic basaltic volcanics in the Snake River Plain had interacted in some fashion with a lake. This indicates that, although the level fluctuated, a large lake existed for a long time.

While the lake existed, basaltic volcanism yielded water-affected basalt, pillows, and tuff cones. After the lake drained but the ground was still saturated with water (high ground water table), the basalt volcanism became explosive and many maars and tuff rings were formed. When the ground dried out like the present, the volcanism became effusive and shield and cinder cones and subaerial lava flows were formed. Finally, stream dissection coupled with the final stage of volcanism produced lavas that flowed into rivers and stream channels, dammed the rivers, and developed small temporary lakes.

Western Snake River Plain: An Outstanding Area to Study Hydrovolcanic Rocks. Based on years of field mapping in the western Snake River Plain, Godchaux and Bonnicksen have reported on a wide variety of hydrovolcanic forms and features in this remarkable area. The western Snake River Plain graben has been the site of Lake Idaho, more or less continuously, from about 12 to 2 million years ago. Large-volume silicic eruptions occurred between 12 and 11 million years ago. Rhyolite-water interactions produced near vent breccias, subaerial phreatomagmatic tuffs, and silicified massive breccias. They suggest that the western Snake River Plain is one of the best places in the world to study the wide variety of volcanic forms that are produced when magma and lava interact with surface water because of its relative youthfulness and excellent exposures,

and the fact that the lake is now drained.

Evidence of Basaltic Volcanism for a Permanent Lake Idaho

Jenks and Bonnicksen (1989) propose that the Lake Idaho shorelines can be defined on the basis of the basalt-water interaction. The best evidence of the ancestral shoreline is the elevation and geographic position where basalt flows pouring into the lake were first affected by water. The 3800-foot elevation contour represents the highest stand of the lake shoreline. The 3800-foot elevation contour also represents a "major regional topographic break" where it delineates many Tertiary lacustrine basins in western Idaho and eastern Oregon.

The Bruneau and Jarbidge Rivers, as well as other streams that emptied into ancestral Lake Idaho, have cut their canyons after the final draining of the lake waters. The widespread gravel layers that overlie the lacustrine sedimentary rocks were deposited from the draining of the ancestral Bruneau River, Salmon Falls Creek, Clover Creek and Sheep Creek into the basin as the lake receded for the final time. These gravels were deposited by braided streams before the downcutting of the canyons occurred and before the final recession of the lake.

Basin-Fill Sedimentary Rocks

The consensus of most geologists working in the western Snake River Plain is that the sediments are primarily of lacustrine origin. There is a particularly diverse depositional environments along the basin margins with floodplain and fluvial deposits common. At the Boise Front, the lithologies, textures, and environments of deposition were controlled by the granitic mountains to the north and the Boise River which drained thousands of square miles of mountainous country for several million years.

Idaho Group Sedimentary Rocks

The sedimentary rocks of the Idaho Group in the northern part of the western Snake River Plain do not correlate well with the Chalk Hills and Glens Ferry Formations in the southern part of the Snake River Plain. Consequently, Idaho Group rocks near the Boise Front were named Pierce Gulch and Terteling Springs Formations.

Most of the Glens Ferry Idaho Group deposits are lacustrine and along the basin margins they are



Pliocene Lake Idaho sediments near Hagerman, Idaho.

influenced by the type of sediments readily available from the highlands. For example, along the Boise Front, the composition textures and depositional environments were controlled by the large granitic source to the north resulting in typical lake margin environments.

Gravel Cap: Tenmile and Tuana Gravel

- *Tuana Gravel.* The Tuana Gravel lies on the south side of the Snake River in the vicinity of King Hill on benches approximately 800 feet above the Snake River. The distribution of these gravels indicates that they are remnants of alluvial fans formed by the ancestral drainage of Salmon Falls Creek on the East and the Bruneau River to the west. The dissected toes of these fans are 50 to 80 feet below the estimated spillover level of Lake Idaho at the head of Hells Canyon.
- *Tenmile Gravel.* The Tenmile Gravel, situated in a number of patches south and southwest of Bois, has an elevation of 2500 feet for its most western deposits; at this point the gravel is about 350 feet above the Snake River. The gravels deposited by high-energy braided streams ended the basin-filling sequence. Once the basin of the western Snake River Plain was filled, perhaps between 1.6 and 1.7 Ma, the major rivers that flowed into the plain began prograding coarse-grained, braided stream channel deposits across the broad, flat surface of the plain. The deposition of the Tenmile Gravel, which overlies lacustrine and flood-plain deposits, indicates an increase in stream discharge and bed load. This was probably caused by higher runoffs of snowmelt and the advance and retreat of glacial ice during the Pleistocene ice age.

Eastern Snake River Plain

The eastern Snake River Plain is a volcanic province with thick sequences of rhyolite flows capped by undissected basalt flows. The basalt overlaps sedimentary deposits of the western Snake River Plain. The eastern plain is 55 to 60 miles wide and has a length of 210 miles. The elevation rises from almost 3280 feet at Hagerman to almost 6560 feet north of Ashton. Although Paleozoic rocks both north and south of the Snake River Plain have similar stratigraphy and structure, there is no evidence that these rocks continue across or underlie the plain.

Downwarp

The eastern Snake River Plain is a downwarp rather than defined by boundary faults. No solid evidence for boundary faults exists. Both ash-flow tuffs and Miocene to Pliocene-age sedimentary rocks dip toward the axis of the plain. Rhyolite rocks in the Snake River Plain are generally thought to originate from a hot spot located in the mantle below the southwestward-moving North American Plate.

Exploratory wells at the INEL indicate the rhyolitic rocks are at least 2 miles thick and seismic surveys indicate that the base of these rocks could be at least 3.5 miles deep. Subsurface plutons, which were the source for the immense ash-flow eruptions may be situated between 2 and 5 miles below the surface. Also, gravity anomalies may indicate blocks of Paleozoic carbonate rocks between the caldera.

The rhyolitic rocks are progressively younger towards the east starting with dates of 11.3 million years at Bruneau Jarbidge eruptive center to about 4.3 million years at the east end of the Snake River Plain. Of course, the most recent rocks on the topographically high Yellowstone Plateau occurred as recently as 0.6 million years ago.

Caldera Complexes

The Island Park and Yellowstone caldera complexes are situated northeast of the eastern plain. The Rexburg caldera complex on the eastern plain has been identified and the existence of many others has been inferred. Silicic volcanic rocks and associated intrusive bodies apparently underlie most of the eastern plain. Rhyolite is far more abundant than basalt in the eastern plain—only a thin veneer of basalt overlies a thick sequence of rhyolitic ash and flow tuffs. It is very likely that many source calderas for rhyolite are buried below the basalts.

Heise Volcanic Field

Ash-flow tuffs of the Heise volcanic field cover 13,500 square miles and ash from the eruptions was distributed over much of the western United States. The three flows of the field, the Blacktail Tuff (6.6 my), the Walcott Tuff (6 my), and the Kilgore Tuff (4.3 my) were extruded from three calderas.

Island Park Caldera

The Island Park Caldera may be the largest symmetrical caldera in the world. Rhyolite was erupted during the initial collapse period. Then basalt and rhyolite were alternately erupted from vents along the caldera floor. Finally basalt was erupted. The tuffs, flows and fault scarps that make up the morphology of the caldera are so young that they have been modified very little by erosion. There were three basic phases in the evolution of the caldera: growth of the volcano, extrusion of magma, and collapse.

The Island Park Caldera is an elliptical collapse structure 18 to 23 miles in diameter and is situated in the center of a rhyolite shield. The western semicircle of the scarp is exposed, and the eastern semicircle is buried under flows of rhyolite. Both basaltic and rhyolitic lava are believed to have originated from a single magma chamber below the caldera.

The rim crest of the southwestern side of the caldera stands about 1200 feet above the plain south

of the caldera. The most abundant rock type composing the caldera is flow tuff, followed by ash falls and lava flows. The central portion of the caldera collapsed along a semicircular zone of faults 18 miles in diameter at the western half of the caldera.

Island Park-Yellowstone Plateau

The Island Park Yellowstone plateau area has experienced catastrophic eruptions of 70 to 600 cubic miles of ash-flow tuff. The Huckleberry Ridge Tuff eruption was the largest and formed about 2.1 million years ago. The Mesa Falls Tuff had a volume of 70 cubic miles. About 600,000 years ago the last eruption occurred from the Yellowstone Caldera and covered 250,000 square miles.

Yellowstone National Park embraces 1400 square miles and has an average elevation of 7,000 feet. The Yellowstone Caldera is the world's largest thermal area and has outstanding examples of hot springs, geysers, and mudpots. The caldera is oval shaped, about 45 miles long and 25 miles wide. The Yellowstone Plateau is seismically very active; 15,000 quakes measuring greater than 2 on the Richter scale were recorded between 1973 and 1988. In 1959 the 7.5 Hebgen Lake Earthquake originated on a fault 10 miles west of Yellowstone Park. The Yellowstone plume or hotspot has produced about 1,000 cubic miles of rhyolitic ash during the last 2 million years.

Yellowstone has about 10,000 hot springs and several hundred geysers. These form when rain and snow melt and migrate to a depth of about a mile through permeable gravels and fractures in the volcanic rock. After heating, the hot steam and water move upland along fractures. Geysers are a type of hot spring that intermittently erupt hot water or steam. They require special conditions such as (1) a source of heat from magma or very hot rocks, (2) abundant water, (3) a permeable fault or fracture system, (4) near-surface underground storage area for water and steam, and (5) a small surface opening to force steam or water out under pressure. Upper Geyser Basin of Yellowstone National Park has the greatest concentration of active geysers in the world.

Snake River Group

The Snake River Group includes most of the basalt flows in the Snake River Plains that were extruded during the Pliocene to Holocene epochs. The youngest flows are no more than a few thousand years old and the oldest are about four million years old. Approximately 8,000 square miles of southern Idaho are covered by basalt flows and interbedded sediments of the Snake River Group.

Geophysical studies and drill holes indicate that the plain may be underlain by basalts as much as five miles thick. However 5,000 feet may be an average thickness. Because only the upper 1500 feet were sampled by drill hole, it is possible that some of the basalts not reached by the drill hole are flows of the Columbia River Group.

The Snake River Basalts tend to be extruded from central vents rather than fissures. In the Snake River Plain there are numerous small shield volcanoes 200 to 400 feet high. On a clear day, from almost any place on the Snake River Plain, one can see low hills on the horizon that are either cinder cones or shield volcanoes of basalt. At Craters of the Moon National Monument, very recent flows and volcanic structures can be examined.

There are eight Holocene basaltic lava fields in the eastern Snake River Plain which are closely related to the volcanic rift zones: (1) Shoshone, (2) Craters of the Moon, (3) Wapi, (4) King's Bowl, (5) North Robbers, (6) South Robbers, (7) Cerro Grande, and (8) Hells Half Acre. Although the older surfaces are covered with loess, in general the surface features are unchanged from emplacement.

Quaternary Basalt

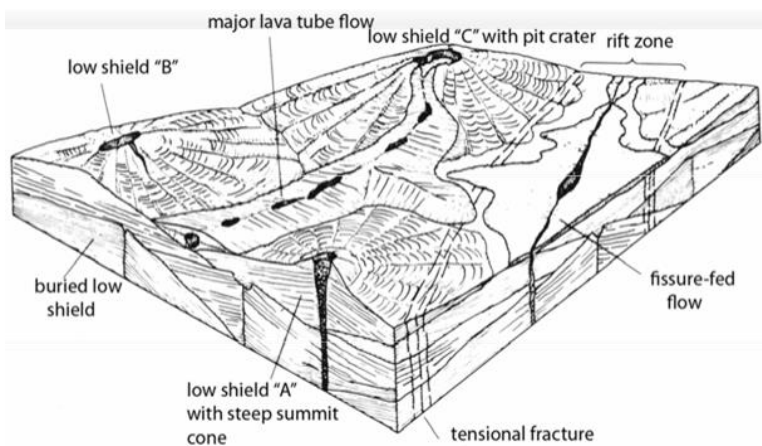
Quaternary basalt eruptions on the plain generally occurred after the plate passed over the hot spot. Potassium-argon dates are generally less than 2.0 million years ago. Geophysical studies (resistivity) suggest that the basalt of the eastern Snake River Plain is generally less than 1-mile thick. As expected, overall the basalt thins towards the edge of the plain and is thickest near the axis. Drilling information on the west side of the eastern Snake River Plain indicates that the basalt thins to the west, and accordingly, so does the water table. Sources of basalt or the vents tend to be concentrated along the axis of the eastern Snake River Plain, forming a subdued medial ridge. Basalt vents are also concentrated along the volcanic rift zones that tend to be aligned at right angles along the rift.

Volcanic Rift Zones

There are at least 8 rift zones in the eastern Snake River Plain that have contributed basalt to the plain. During the Holocene, basaltic eruptions were scattered across the entire eastern plain. Volcanic rift zones may be extension of range-front faults to the north of the plain. The extension indicated by the rift zones may also be related to the basin and range structures both north and south. However, these volcanic rifts may also indicate thermal contraction.

Basaltic Lava Flows

Most basalt flows in the Snake River Plains are pahoehoe basalts that were emplaced as compound flows. Compound flows are a sequence of thin individual cooling units ranging from less than 3 feet to more than 30 feet thick. The



Block diagram showing shield volcanoes, lava tube flows and fissure flows (from Greeley, 1981)

surface is hummocky with local relief, typically less than 30 feet. Common features are tumuli, pressure plateaus, pressure ridges, flow ridges and collapse depressions. These features are caused by the way in which flows advance through a series of budding pahoehoe toes. Low areas and swales are filled with wind-blown sediments so that now, on the surface of older flows, only the higher ridges are visible. Aa flows are not as extensive as pahoehoe flows and can best be seen at Craters of the Moon. Basaltic lava is extruded on the Snake River Plain in three ways: flows from a central vent forming low shields, fissure flows and tube-fed flows.

Low Shields

Low shields are characterized by a small size and low profile. They have slope angles of 0.5 degrees and average 10 miles across and less than 1.6 cubic miles of lava. An excellent example is the Wapi lava field. This field covers 116 square miles with compound lava flows of pahoehoe. It is characterized by features such as lava toes, collapse depressions, flow ridges and pressure ridges, but lacks lava tubes. The Wapi lavas have a carbon 14 date of 2,270 years. Many of the low shields have pit craters at the summit and many craters show evidence of collapse. Shields tend to be aligned along rifts or fissures.

Fissure Flows

Fissure vents are associated with rift zones. The youngest fissure flows on the Snake River Plain are located at Kings Bowl and the Craters of the Moon National Monument. Craters of the Moon covers 580 square miles whereas Kings Bowl covers about 1 square mile. Flow thickness is generally less than 5 feet although ponding or lava lakes may cause thicker layers. Point source eruptions along a fissure are common. For example, spatter cones and cinder cones at Craters of the Moon are examples of point source eruptions.

Lava Tube Flows

Lava tubes and channels originating from fissures and low shields are a very major conveyance for the emplacement of lava. These tubes are commonly more than 12 miles long and range from 2 to 30 feet across.

Cinder cones are not as common in the eastern plain as the western plain. They are generally found along the rift zones.

Shield Volcanoes

Shield volcanoes cause most of the basaltic landforms on the central and eastern Snake River Plain and represent most of the total volume of basalt. There may be 8,000 shield volcanoes on the eastern Snake River Plain. Lava may flow as much as 80 miles from the source vent and may cover an area up to 145 square miles. Shield volcanoes have very gentle slopes on the flanks of 1 to 2 degrees with an overall height of 150 to 450 feet above their base. The lower flanks tend to be covered by overlapping flows from adjacent volcanoes. The volcanic eruptions have been

episodic rather than periodic with intervals between eruptions ranging from less than a few thousand years to more than a few hundred thousand years.

Common features on the flows are pressure plateaus, pressure ridges, flow ridges and collapse depressions. These features are caused by the way in which flows advance through a series of budding pahoehoe toes. Low areas and swales are filled with wind-blown sediments so that now, on the surface of older flows, only the higher ridges are visible. Aa flows are less common and tend to be thicker than pahoehoe flows. As a general rule aa flows in the eastern Snake River Plain occur where the SiO₂ content of the lava is greater than 50 percent; an exception to this rule may occur when the lava descends a steep slope.

Black Butte

The Black Butte or Shoshone Lava Field covers approximately 71,000 square miles and based on a radiocarbon date is 10,130 years old. Black Butte is such a low shield volcano that it is barely noticeable as a volcano from highway 75. The cone rises approximately 200 feet above the surrounding lavas and has a 2-mile basal diameter. In the center of the shield is an irregularly shaped, 80 to 100-foot-deep subsidence crater approximately 0.6 mile in its largest diameter. Subsidence of what was once a lava lake occurred



Aerial view of Black Butte, a shield volcano in southcentral Idaho. The cone rises approximately 66 m above the surrounding lava plain and has a 3.2 km basal diameter (Kuntz, 1986). In the center of the shield is an irregularly shaped, 26 to 66 m deep subsidence crater approximately 1 km in its largest diameter. The crater was once occupied by a lava lake which overflowed the rim and washed thin sheets of foamy pahoehoe down the flanks. The lake eventually drained through a south-southeast trending lava tube.

as the molten lavas flowed and drained through a south-southeast trending lava tube which is now largely collapsed. When the lake was at its highest level, waves of molten lava washed over the crater rim depositing sheets of foamy pahoehoe on the flanks. Some of this pahoehoe was converted to aa lava as the sheets were contorted during final crystallization. The Black Butte flow exhibits very little weathering because vegetation is sparse and the basaltic lava is still dark and crystalline in appearance.

Shoshone Lava Cave System

Lava tubes are important for the emplacement of lava on the Snake River Plain. Lava tubes are the subsurface passage ways that transport lava from a vent to the site of emplacement. They form only in the fluid pahoehoe flows. Tubes originate from open-flow channels that become

roofed over with crusted or congealed lava. However, a tube may also form in a massive flow. Lava tubes exist as a single tunnel or as complex networks of horizontally anastomosing tubes and may occupy up to five levels. Most tubes tend to be 6 to 15 feet across. Access to some tubes may be gained through collapsed sections. Features in tubes include glazed lava, lava stalactites and ice.

Because lava tubes and channels originating from fissures and low shields are a very major conveyance for the emplacement of lava, they are commonly more than 12-miles long and range from 3 to 30-feet across. Shoshone Ice Cave is a segment of a complex lava tube and lava channel system, i.e. a system that had both roofed and unroofed segments. Many of the roofed parts later collapsed. The Shoshone lava tube system covers approximately 80 square miles, with smaller subsidiary tubes branching off from the main tube allowing lava removal from the main tube.

Rhyolite Domes

Four Rhyolite domes aligned parallel to the axis of the plain penetrate the Holocene basalts of the eastern Snake River Plain: Big Southern Butte (0.304 my) East Butte (0.58 my), Middle Butte, and an unnamed butte (1.42 my).

Big Southern Butte, Middle Butte and East Butte are three large buttes which can be seen rising above the eastern Snake River



Big Southern Butte, a rhyolite dome, rises 2500 feet above the plain. It was an important landmark for the early settlers.

Plain while driving between Arco and Idaho Falls on Highway 20. All three buttes are situated east of the highway. Middle Butte appears to be an uplifted block of basalt. Although no rhyolite is exposed at the surface, the butte was probably formed by a silicic intrusion forcing the basalt upwards into the form of a butte. Big Southern Butte and East Butte are rhyolite domes. East Butte has been dated at 600,000 years, whereas Big Southern Butte has been dated at 300,000 years.

Big Southern Butte, because of its prominence and size, was an important landmark for the early settlers. It rises 2,500 feet above the plain and is approximately 2,500 feet across the base. Access to the top of the butte is available by a Bureau of Land Management service road. The butte was formed by two coalesced domes of rhyolite that uplifted a 350-foot section of basalt. The basalt section now covers most of the northern side of the butte. The dome on the southeastern side was developed by internal expansion (endogenous growth). Rupture of the

crust at the surface caused breccia to form. Obsidian, pumice and flow-banded rhyolite are important components.

Subsidence of the Plain

Subsidence of the plain southwest of the hot spot occurs from thermal contraction after the plate passes by the hot spot. This thermal contraction also causes the tuffs at the margins of the plain to dip towards the axis.

Drainage

No drainage system has been established on the surface of the basalt. Most of the precipitation that does not evaporate, seeps into the basalt and enters the Snake River Plain regional aquifer system.

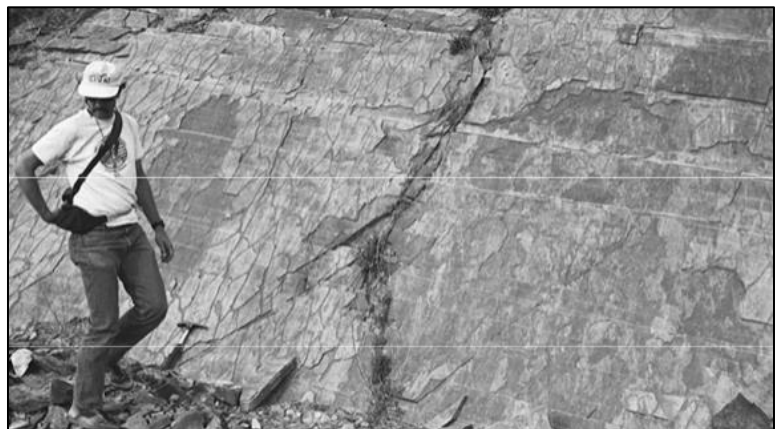
Soils

Most of the wind-blown silt (loess) transported by wind storms is derived from detrital sedimentary material in lakebeds. This soil is not derived from the underlying rock. Where the winds are sufficiently strong to carry in sand-sized material, dunes may form and cover the volcanic surface.

Pre-Cenozoic Sedimentary Rocks of Eastern Idaho

Precambrian Rocks of Eastern Idaho

Paul Link has mapped and described the Pocatello Formation as the oldest rocks in southeast Idaho. These rocks, which were deposited 750 to 700 million years ago, consist of two rock types: metamorphosed basalt and a diamictite consisting of pebble- to boulder-sized material. The diamictite consists partly of glacial till representing a Precambrian ice age in Idaho. During this time, North America was connected to Antarctica or Australia and Idaho was much closer to the equator.



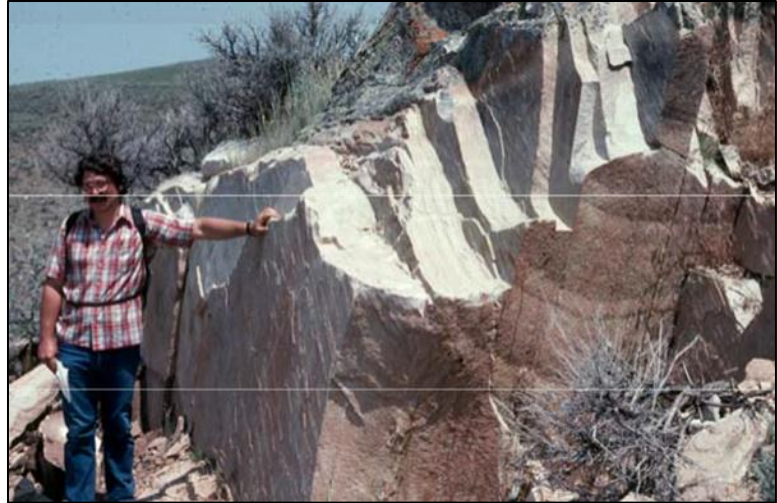
Exposure of the Precambrian Yellowjacket Formation, an argillaceous siltite. The mudcracks, which indicate shallow-water deposition, are elongated down dip on the steeply dipping bedding surface. Ellis, Idaho.

The Brigham Group, which overlies the Pocatello Formation, was deposited from 700 to 530 million years ago. Many of these rocks which were deposited as sandstones in shallow oceans

have since been metamorphosed into quartzite.

Paleozoic Rocks of Eastern Idaho

After the Brigham Group, a thick sequence of limestones was deposited on broad carbonate platforms in the warm shallow seas along the continental margin of western North America. These Paleozoic limestones contain a variety of invertebrate fossils such as Bryozoans, corals, mollusks, trilobites, and brachiopods. Thin units of fine-grained sandstone are interbedded within the Paleozoic limestone. Some of the limestone has been altered to dolomite.



Steeply dipping beds of the resistant Nugget Sandstone Formation east of Bear Lake in eastern Idaho. These rocks were formed from Jurassic sand dunes.

Paleozoic Rocks of East-Central Idaho

In the Pioneer, Boulder, and Smoky Mountains near Sun Valley there are thick sequences of black shales and argillites known as the Devonian Milligen Formation. More than 15,000 feet of Mississippian conglomerates and sandstones of the Copper Basin Formation were deposited in a deep-sea fan environment. These rocks are primarily exposed in the Southern Pioneer Mountains. Over in the Wood River area, the Milligan Formation is overlain by thick calcareous siltstones and sandstones of the Upper Pennsylvanian and Permian Sun Valley Group. These sediments were deposited in deep water west of a carbonate bank.

Mesozoic Rocks of Eastern Idaho

Mesozoic-age rocks in southeast Idaho include thick sequences of limestones, sandstones, and shales. These sediments were deposited in shallow basins caused by the weight of the advancing thrust sheets.

Idaho-Wyoming Thrust Belt

The Idaho-Wyoming Thrust Belt is located in southeastern Idaho, western Wyoming and north-central Utah. This thrust belt is a small part of a much larger, geologically-complex Overthrust Belt that extends 2200 miles from Alaska into Central America. The belt is a part of the extensive north-south-trending North American Cordillera Mountain Range which originated by intense east-west compressional deformation spanning more than 200 million years. Many

different thrust sheets comprise the belt. These sheets were typically transported eastward for distances of 10 to 20 miles. Total displacements may exceed 70 miles and total shortening was about 70 percent.

Cause of Thrusting

During the accretion process, the eastward-moving Pacific Plate was subducting or descending along a deep oceanic trench beneath the western continental margin of North America. This subducting plate caused the formation of numerous magma bodies to move upwards into the continental crust and form the great Sierra Nevada and Idaho Batholiths. Possibly in response to the upward intrusion of these large granitic masses, the Overthrust Belt formed to the east. The thrusting was driven by gravity and resulted from crustal thickening caused by emplacement of the batholith. This uplift of the Idaho Batholith was accompanied by the development of folds and thrust faults as well as possible gravitational gliding off the highlands.

Date of Thrusting

The Paleozoic and Mesozoic sediments were strongly folded and thrust eastward beginning during Late Jurassic (about 120 million years ago) in the west and ending as late as Eocene time (52 million years ago) in the east. This 70 million year period of compressive tectonics was followed by normal or extensional faulting of the Basin and Range within the Overthrust Belt. Therefore, the ages of foreland thrusting are determined to be youngest in the east and oldest in the west.

Sevier Orogeny

The period of intensive Late Jurassic to Eocene compressive thrust faulting and folding in the Cordilleran Overthrust Belt has been designated the Sevier Orogeny. During this period, huge masses of rock were ruptured into thrust sheets 20 to 50 miles wide, several miles thick and hundreds of miles long. Eastward horizontal movement on the individual thrust sheets is measured in tens of miles.

The Sevier Orogenic Belt is a narrow north-south zone of structural disturbance that runs through western Montana, eastern Idaho, Wyoming, Utah and Nevada. This belt consists of large-scale thrust faults and folds with vergence to the east. These thrust faults and folds have resulted in significant shortening of about 70 percent that was initiated in Late Jurassic time and continued through the beginning of the Tertiary.

On the basis of fossil distribution found in the earliest conglomerate deposited during the orogeny within the thrust belt, the initial thrust movement may not be older than 119 million years. The rapid increase in subsidence of sedimentary sequence within and east of the thrust belt is believed to have been caused by the initiation of faulting in the thrust belt.

Four Major Thrust Fault Systems

Four major thrust-fault systems have been identified in the region. From west to east and oldest to youngest: Paris-Willard, Meade-Laketown, Crawford, Absaroka, and Darby. Each major thrust fault has many overlapping thrust slices. Before horizontal compression, the Overthrust Belt was approximately 130 miles wide; however, total stratal shortening of 70 percent was accomplished by thrust faulting and folding. The thrust plates, which are 10,000 to 20,000 feet thick, are ruptured along flat faults more or less parallel to bedding in weak or incompetent rock layers (such as shales), and cut steeply across bedding in strong or competent layers.

Younger Rocks Thrust over Older Rocks

Only the sedimentary rock section is shortened by folding and faulting. The Paleozoic and Mesozoic rocks are structurally detached from the Precambrian granitic basement rocks below by a regional decollement (detachment surface) and shifted eastward over the Precambrian basement rocks along the decollement or thrust surface. Despite the compressive deformation to the overlying Paleozoic and Mesozoic rocks, the Precambrian basement rocks are little affected. As a result of the thrusting and imbrication of strata, older rocks are thrust over younger rocks. In some cases, early Paleozoic rocks in the hanging wall are thrust over Cretaceous rocks.

Folds

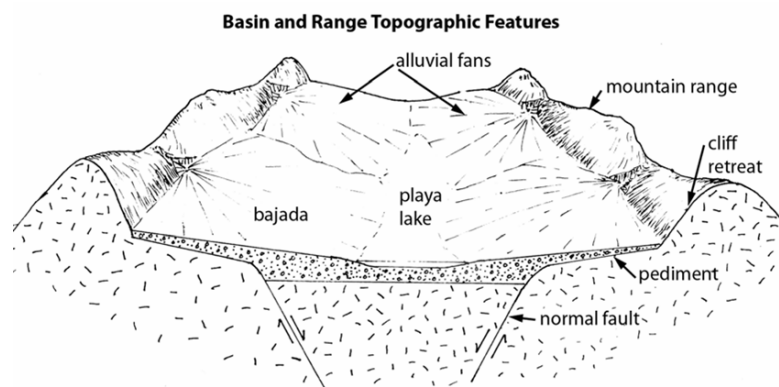
Like thrust faults, folds for the most part affect only strata above the decollement surfaces. Folds are generally concentric in form and are tilted to the east. Folds have amplitudes of 5,000 to 10,000 feet.

Extension through Snake River Plain

The Cenozoic Snake River Plain effectively divides the thrust belt. Correlation of geology north and south of the plain is difficult. However, evidence indicates that at one time the thrust belt was continuous before the plain existed. There is no evidence that overthrust structures underlie the Snake River Plain.

Basin and Range Province

Basin and range structure is caused by uplifted block-faulted mountain ranges separated by valleys. The physiography of the States of Arizona, New Mexico, California, Utah, Nevada, southern Oregon and southern Idaho is dominated by these long, linear mountain ranges



Generalized cross section shows how crustal extension causes several different types of normal faults resulting in Basin and Range topography.

and intervening valleys. The valleys and mountains are separated by high-angle extension faults. The valley elevations range from 4,200 to 5,200 feet, whereas the mountain crests range from 6,500 to 11,800 feet.

The Basin and Range Province is characterized by mountain ranges separated by flat valley floors. Each range is bounded by large faults along which the range has moved upwards relative to the adjoining basins or valleys. In other words the topography is strongly controlled by faulting.

Bajada

Heavy rains promote rapid erosion along the steep mountain fronts. Loose rock and sediment are washed down narrow stream channels and are deposited as alluvial fans at the base of the ranges. The water from the stream may flow to the valley bottom and form a playa lake. Playa lakes are shallow and temporary, lasting only several days after a storm. Continued deposition of alluvial fans at the base of a mountain may form a bajada. A bajada is a broad, gently-sloping depositional surface formed by the coalescing of alluvial fans. A bajada may have a gently rolling surface caused by the merged cone-shaped fans.

Pediment

A gently-sloping, flat erosional surface at the foot of the mountain is called a pediment. This surface is generally covered with a veneer of gravel. The pediment surface develops uphill from a bajada as the mountain front retreats. As you might suspect, it is difficult to distinguish a pediment surface from a bajada surface because both have the same slope and gravel cover. However, the pediment is an erosional surface, whereas the bajada is a depositional surface and may be hundreds of feet thick.

Normal Faults

There is a complex system of normal faults in the Basin and Range that controls the topography. The mountains are 9 to 12 miles across and they alternate with alluviated valleys of approximately the same width. The ranges or linear mountain blocks are formed by vertical movements along faults on one or both sides of the block. Faults are also distributed throughout the ranges and valleys.

The mountains (horsts) and valleys (grabens) are bounded on both sides by normal faults. The mountain ranges tilt from a few degrees to more than 30 degrees. On the east side of the province, tilting tends to be towards the east; whereas, on the west side, tilting tends to be towards the west.

Valley fill ranges from about 500 feet to more than 9,800 feet thick. The structural relief between bedrock in the valleys to the crest of the adjacent mountains ranges from 6,500 to 16,000 feet.

Therefore, the faults have up to three miles of vertical displacement.

The ranges may be tilted by asymmetric grabens, rotation of blocks along downward flattening faults (listric faults), and rotation of buoyant blocks.

Crustal Extension

The normal faults of the Basin and Range are indicative of extension normal to the fault lines. The amount of extension depends on whether the faults flatten at depth or continue at the dip of 60 degrees. If the faults do not flatten and continue down to where brittle rupture is replaced by laminar flow, the extension would only be about 10 percent. But if the faults flatten downward, extension might be as much as 50 percent of the present width. Tilting of Tertiary welded tuffs and lake bed sediments that were horizontal to sub-horizontal before faulting require rotation tilting by the faulting. This can only be explained by flattening downward of major faults.

The crust of the southern Basin and Range terrane is about 15 miles thick; whereas, the crust of the Colorado Plateau is about 25 to 28 miles thick. This thin crust may be a function of the amount of extension in the Basin and Range region. Therefore, on this basis, approximately 40 percent of the present southern Basin and Range may represent extension. Using the same approach, 30 percent of the present width of the northern Basin and Range may represent extension.

Further evidence of crustal extension includes high surface heat flow, voluminous magmatism, low velocity of seismic waves in the upper mantle, and high attenuation of seismic waves in the upper mantle. In the northern Basin and Range, the structures continue across the Snake River Plain but not the Idaho Batholith.

Age of the Basin and Range

The late Cenozoic structure that formed the Basin and Range Province began about 17 million years ago. About 10 million years ago the present topography was formed.

Relative Vertical Displacement

Gravity studies in the Lemhi Birch Creek Trench indicate a structural relief of about 13,100 feet from the crest of the Lemhi Range to the base of the Tertiary fill in the valley bottoms. Therefore, the relative vertical displacement along the steep reverse faults is about 13,100 feet.

Bonneville Flood

Lake Bonneville, a late Pleistocene lake, was the precursor of the Great Salt Lake. The shorelines of this ancient lake can be seen on the higher slopes of the Wasatch Mountains, more than 1060 feet above the present level of the Great Salt Lake. Before the catastrophic Bonneville flood,

Lake Bonneville covered an area of more than 19,691 square miles.

Approximately 14,400 years ago Lake Bonneville suddenly discharged an immense volume of water to the north. This flood is thought to be caused by capture of the Bear River which greatly increased the supply of water to the Bonneville Basin. The release of water from Lake Bonneville was apparently initiated by sudden erosion of unconsolidated material on the northern shoreline near Red Rock Pass. These flood waters flowed over Red Rock Pass in southeastern Idaho and continued westward across the Snake River Plain generally following the path of the present Snake River. Most of the important depositional and erosional features of the Bonneville Flood were developed in a few days; however, the Snake River sustained a high rate of flow for more than a year.



Looking east at an enormous pothole enlarged by the Bonneville Flood. The flood also excavated a large amount of basalt north and west of the pothole. The pothole is situated at the foot of Shoshone Falls.

Gilbert (1878) discovered the evidence for the flood during field study of Pleistocene Lake Bonneville between 1876 and 1879. He recognized many of the catastrophic effects of the flood between Red Rock Pass and Pocatello, including erosional features 400 feet above the Marsh Creek flood plain. Although this enormous flood was first described in the literature by Gilbert in 1878, Harold Malde (1968) of the U.S. Geological Survey published the first detailed account of the effects of the flood on the Snake River Plain. The name "Bonneville Flood" first appeared in the literature in 1965.



Looking northeast at alcoves along the north rim of the Snake River Canyon. Note the scabland on the rim where the flood overflowed the canyon and swept across the Rupert channel, poured back into the canyon, and formed the alcoves by cataract retreat.

Colossal Erosion near Twin Falls

Partly because of the reentry of the upland channel, the Snake River Canyon in the vicinity of Twin Falls experienced the most exceptional bedrock erosion of any place on the Snake River canyon. This highly eroded canyon segment is more than a mile wide in places, 500-feet deep, and more than 10-miles long. The most extraordinary features are the huge spillway alcoves on the north rim of the canyon where the water from the overflow channel reentered the canyon. A series of large northwest-trending alcoves on the north rim are approximately parallel to the Rupert channel. These alcoves and the associated scabland on the north rim represent a 10 mile-wide spillway. Other remarkable features include the "underfit" Shoshone Falls. One fourth to one-third of a cubic mile of basalt was ripped from the canyon in the Twin Falls area and carried downstream.

The Shoshone Falls are a remarkable erosional remnant of the Bonneville Flood. The falls, as well as all the associated spectacular erosional features in the 16 km-segment of the Snake River canyon, were created by the eight-week-long flood. Although the falls are a spectacular 212-foot-high feature, they are framed by diverse, large scale flood-caused features such as cataracts, spillway alcoves, and scabland. These features apparently occur at the convergence of two channels of the Bonneville flood—the Rupert channel and the Snake River channel. This 10-mile-long segment of the north rim also offers the most spectacular erosional effects of the flood and is somewhat analogous in origin and form to the Dry Falls cataract complex in central Washington. The largest alcoves, from east to west, are Box Canyon, Devils Corral, and Blue Lakes Alcove. Blue Lakes Alcove and Devils Corral are enormous features, almost one mile in length and more than a quarter mile in width.

On the south rim of the canyon, two large south-southeast-trending alcoves and a ragged rim offer evidence that the overflow waters also reentered the Snake River over the south rim. The erosion and the development of these alcoves on the north and south rim did not occur by water cascading over the rims because the canyon was undoubtedly full during most of the erosion. Most of the erosion must have been accomplished by extraordinary currents in flood waters more than 500-feet deep.

Scabland

J. Harlan Bretz coined the word "scabland" to describe the stripped basalt surface in southeastern Washington. The term, now well established in the geological literature refers to not only the bare basalt which had the loess swept off by the flood, but it also covers a variety of gigantic erosional features associated with the flood such as anastomosing channels, coulees, dry falls and empty-rock basins. Similar features, both in size and form, exist along the path of the Bonneville Flood, especially where the flood overflowed the canyon in the vicinity of Twin Falls. All loess has been stripped of the basalt surface along the flood path. If you have the opportunity to fly along the Snake River Canyon following the flood path you can clearly see that there is no

farming within the area swept by the flood waters because all soil was removed down to bare bedrock. In most areas, the agricultural lands come right to the edge of the flood-stripped area and, in effect, define the flood boundaries. In some areas more intense erosion by the flood stripped off the upper few feet of vesicular basalt and formed a butte-and-basin scabland.

Melon Gravels

Large rounded boulders of basalt characterize many deposits left by the flood along the Snake River Plain. H. A. Powers, who recognized that these boulders were of catastrophic origin, and Malde applied the name of “Melon Gravel” to the boulder deposits. They were inspired to use this term after observing a road sign in 1955 that called the boulders "petrified watermelons."



Basalt boulders on a lava flow more than 80 m above the Snake River near Bancroft Springs, Idaho. The long diameter of the largest boulders exceeds 5 m. Note the imbricate structure of the boulders indicating current flow from right to left.

The melon gravel, which consists of rounded basaltic pebbles, cobbles and boulders, is easy to recognize and is good evidence of the flood path. The deposits are characterized by very poor sorting, crude, thick sets of inclined bedding and no horizontal bedding. The average bar is almost 1.5-miles long, 0.5 mile wide, and may be as thick as 300 feet. The bars appear to be preserved intact from their time of deposition showing no perceptible erosion.

King Hill Basin

The King Hill basin contains almost one-third of a cubic mile of melon gravel. This represents more than half of the flood debris deposited in the canyon downstream from Twin Falls. The reason for the occurrence of this immense amount of gravel is the King Hill basin offered the first slack water since the flood waters left Twin Falls. One gravel bar covers 6 square miles and is 160-feet thick. Immense boulders up to 15 feet in diameter were carried 240 feet above the Snake River near Bancroft Springs. These are the largest boulders found downstream from Twin Falls.

The Spokane/Missoula Flood

In 1923, J. Harlon Bretz first proposed that certain erosional features on the Columbia Plateau were caused by the great "Spokane Flood." Though many geologists at the time scoffed at Bretz's proposition, it is now widely accepted that this flood not only occurred some 12,000 to 16,000 years ago during the great ice age, but that it is one of the greatest floods ever recorded by man.

Bretz's involvement with the Spokane Flood was initiated in 1923 with his first published paper on the subject which appeared in the Journal of Geology and culminated with a final publication in the same journal in 1969—his many publications on the subject spanned a remarkable 46 years.

The area called the "channeled scablands" is an oval-shaped area of about 15,000 square miles in southeastern Washington. The bedrock of this area is composed of the extensive flows of the Columbia River Basalt erupted during the Miocene Epoch, between 30 million and 10 million years ago. This area has low relief, is situated at an elevation of 2,500 feet, and is surrounded by mountains. After the last flow of lava, a blanket of wind-blown silt or loess accumulated on the lava field. Where these silts still exist west of Moscow, they make up the very fertile soils of the Palouse country of southeastern Washington. This mantle of loess ranges in thickness from less than a foot to several hundred feet and forms many low hills.

The story of the Spokane Flood began about 100,000 years ago when continental glaciers were moving southwest from the great ice fields in British Columbia. The Purcell lobe of the ice sheet moved southward into the Purcell Trench and plugged the Clark Fork Valley. The ice dammed the water of the Clark Fork River near the place where it runs into the Pend Oreille Lake. The impounded water filled many tributary valleys to the east and formed the largest lake in the Pacific Northwest during the great ice age.

Glacial Lake Missoula

Glacial Lake Missoula in western Montana covered approximately 3,000 square miles and had an estimated 500 cubic miles of water. The lake was about 2,000 feet deep near the ice dam and about 950 feet deep at Missoula. Prominent wave-cut shorelines of glacial Lake Missoula can be easily observed on Sentinel Mountain from the city of Missoula. Wave-cut shorelines are generally not well developed indicating the lake did not remain at any one level for a long time. The successive terraces indicate a gradual filling of the lake. Meltwater from both alpine glaciers and the continental ice sheet fed the lake and raised the lake level.

Release of Flood Waters

When the lake level reached the top of the ice dam it is probable that little time passed before the entire dam was breached. An overflowing stream rapidly cut down through the ice and increased the volume of water and the size of the channel. The ice dam could have been breached and the dam destroyed within a day or two of the first overflow.

When the flood waters were suddenly released, the immense amount of water ran south and southwest out of the mouth of the Clark Fork Valley, through Pend Oreille Lake across Rathdrum Prairie, and down the Spokane Valley. Current velocities are calculated to have reached 45 miles per hour in the narrow parts of the Clark Fork Canyon. Calculations also

indicate that the maximum rate of flow was 9.5 cubic miles per hour or 386 million cubic feet per second. This rate of flow represents about 10 times the combined flow of all rivers of the world.

Giant Ripple Marks

Giant ripple marks can now be found in many places swept by flood waters. These ripple marks are so large that their pattern and shape cannot be detected on the ground; however, aerial photography, which has provided convincing evidence of many flood-caused erosional features, also helped identify the giant ripple marks. The best examples of these ripple marks can be seen on the south side of Markle Pass just north of Perma, Montana. These ripple marks cover a 6-square-mile area and have a relief of 20 to 30



Oblique aerial view of giant ripple marks with barn and cottonwood trees for scale (photograph by P. Weiss, 1970; U.S. Geological Survey Photographic Library).

feet. The individual ridges are approximately 2 miles long and 200 to 300 feet apart. Compare these immense features with standard ripple marks one might see in a stream bed or lake shore with ridges measuring less than 1 inch high and separated by several inches.

Channeled Scablands

The large oval-shaped area in southeastern Washington carved by the flood is essentially a large, flat lava field mantled by loess and slightly tilted to the southwest. When the flood moved over the lava field, the huge volume of turbulent water stripped away several hundred feet of loess down to bedrock and carried off blocks of basalt the size of a truck.

The flood carved immense erosional features in the surface of the plain. Canyons more than 200-feet deep and running for many miles were ripped out of the basalt. Plunge pools, cataracts and many other unusual erosional features formed during the flood.

Most of the flood water swept over the lava field in three major rivers: the eastern-most river was up to 20 miles wide and 600 feet deep; the middle channel was approximately 14 miles wide; and the western most and largest river carved the Grand Coulee which measured 50 miles long and 900 feet deep. The Grand Coulee was eroded by a process called cataract retreat. In this process water runs over a cliff into a plunge pool where the turbulent swirling water erodes the

rock at the base of the falls and undercuts the rock on the upstream side. This undercutting causes the overhanging wall to cave and collapse thus moving the falls and plunge pool continuously upstream.

Jointing in the basalt greatly enhanced the ability of the flood waters to pluck and remove the basalt. Therefore, as erosion progressed upstream, a series of falls and plunge pools were developed along the way. Dry Falls was the last falls to form at the end of the flood.

Through Columbia River Gorge to Willamette Valley

All water from this lake was forced to pass through the Wallula Gap and then westward down the Columbia River Gorge to the Willamette Valley. In the Willamette Valley, remnants of the flood waters formed a lake 400 feet deep. When the lake level quickly dropped, large ice bergs rafted from Lake Missoula were emplaced on the shoreline. When the ice melted, boulders were deposited which can still be seen on this ancestral shoreline.

The Spokane Flood covered 550 miles in its traverse across three states. An estimated 500 cubic miles of water was released from glacial Lake Missoula; the water then crossed northern Idaho on its journey to the Willamette Valley in Western Oregon. Duration of the flood, starting when water was first released from Lake Missoula to the time streams in the flood path returned to normal, is estimated to have been about four weeks.

Proposal for Numerous Floods

Waitt and Johnston (1985) have proposed that glacial Lake Missoula periodically discharged numerous colossal jokulhlaups (glacier-outburst floods). They published detailed evidence demonstrating that the erosional and depositional features of the Columbia Plateau were not caused by a single flood but rather 40, or more, huge floods. In fact, their evidence suggests that the number may be close to 100 floods. On the basis of radiometric dates of ash deposits and shell material interbedded with the rhythmites, it is estimated that Lake Missoula existed for 2,000 to 2,500 years between 15,300 and 12,700 years B.P. (before present).

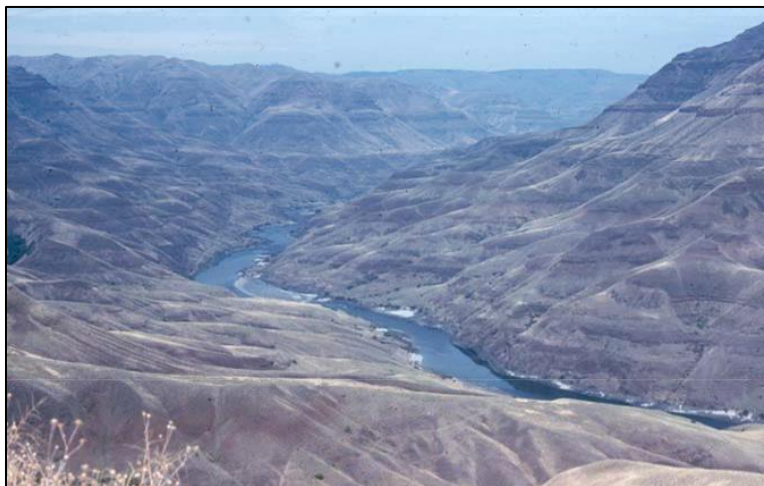
Mechanism for Flood

Waitt and Johnston (1985) also gave a convincing explanation as to how each flood occurred. Lake water did not rise to a level allowing it to spill over or around the ice dam. Before the water reached such a level, the ice dam became buoyant and the glacier bed at the seal broke, causing underflow from the lake below the glacial dam. Subglacial tunnels expanded rapidly and a short time later catastrophic discharge occurred. Calculation of the water budget for glacial Lake Missoula indicates that the lake filled every 30 to 70 years.

Geologic Attractions

Hells Canyon

The portion of the Snake River Canyon that is situated along the boundary between Idaho and Oregon has been referred to as “Hells Canyon,” “Grand Canyon of the Snake,” and “Box Canyon.” Tracey Vallier, a retired U.S. Geological Survey geologist who spent more than 40 years mapping the canyon, considers Hells Canyon to be that stretch of the Snake River Canyon between the Oxbow and the mouth of the Grande Ronde River.



Looking north at Hells Canyon. Note the numerous flow of flat lying Columbia River Basalt exposed along the canyon walls.

So how does the depth of Hells Canyon compare with other canyons in North America? If the depth is measured only from the Idaho side, one segment is more than 8,000-feet deep. If measured from a place near Hat Point, the canyon is about 100 feet deeper than the Grand Canyon of the Colorado. Several canyons in North America may be deeper, including one segment of the Salmon River in Idaho and Kings River Canyon in California.

At Farewell Bend, Oregon, the Snake River Plain of Idaho flows northward 160 miles to the mouth of the Grande Ronde River. To the west are the Wallowa Mountains of Oregon and to the east are the Seven Devils and Cuddy Mountains of Idaho. For most of this distance, the canyon is less than 3,000 feet deep. However, for the 60 mile stretch north of Oxbow Dam the river is 4,000 to more than 5,000 feet below the canyon’s western rim.

The Snake River has flowed through Hells Canyon for at least 2 million years and possibly as long as 6 million years. In 1954, Wheeler and Cook published a paper proposing that the portion of the Snake River north of the Oxbow was once a tributary of the Salmon River. As the river eroded southward, it captured Lake Idaho. The Snake River then cut a channel deeper than the Salmon River, and once that happened, the Salmon River became a tributary of the Snake River. Consequently, the vast amount of water contained in Lake Idaho became the primary agent in forming the great depths of Hells Canyon.

Rocks Exposed in Hells Canyon

The dark rocks, exposed along the lower walls of the canyon, are the old chain of volcanic islands (island arc) that accreted against the continent 93 to 120 million years ago. Gray limestone was deposited on the submerged platform while there was no volcanism. Granitic plutons are exposed, particularly in the Wallowa and Seven Devils Mountains. The youngest rocks exposed in the upper walls of the canyon are the Columbia River Basalt flows. These flows filled in the topography and built a plateau between 17 and 6 million years ago. Because the youngest flows are 6 million years old, we know Hells Canyon cannot be older than the age of the youngest flows.

Boise Valley Terraces

Boise Valley Terraces

Curt Othberg (1994) has described the development of Boise Valley's outstanding record of well-preserved terraces during most of the Pleistocene Epoch. Eight well-developed terraces record successively lower and younger river-cut surfaces covered with several meters of coarse gravel deposits; they were formed by braided-channel deposition during periods of high river runoff. The downcutting resulting in each river-cut terrace was followed by deposition of the gravels. The record of successive terrace development is best preserved southwest of Boise where the sequence in order of decreasing age is the Tenmile terrace, the Amity terrace, the Fivemile Creek surface, the Lucky Peak surface, the Gowen terrace, the Sunrise terrace, the Whitney terrace, the Boise terrace and the bottomland of the Boise River. The terraces formed in response to climate changes instead of base level changes. Most compelling of this evidence is that in the development of each terrace an erosive event is followed by a depositional event and there appears to be periodicity to the terrace sequence.

The succession of terraces in the Boise Valley were developed throughout the Pleistocene and the age of terrace sequence is based on the following evidence: (1) Boise Valley basalt flows are dated between 100 ka and 1.6 million years based on potassium/argon and argon-40/argon-39; magnetic polarities; (3) degree of soil development; and (4) stratigraphic relationships. This terrace development may have started after the capture and diversion of the Snake River drainage (ancestral Lake Idaho) through Hells Canyon sometime between 1.67 and 1.87 million years ago. Rapid downcutting by the rivers in the western Snake River Plain interspersed with periods of coarse gravel deposition led to the sequence of terrace development. The terraces in the Boise Valley may correlate with certain glaciations in the northwestern United States.

Pleistocene alluvial fans were deposited in the Boise Valley during cool periods of mountain glaciation when much greater runoff from snow melt made it possible for large amounts of gravels to be transported out of the mountains. The Boise River originates in and drains a large area in the Sawtooth and Smoky Mountains of central Idaho. These uplands were extensively

shaped by alpine glaciation throughout most of the Pleistocene. In the vicinity of Boise, the river is incised about 500 feet, with a well-defined sequence of terraces south of the westward-flowing Boise River. The Tenmile Gravel represents the basin-filling sedimentation in the Snake River Plain of a late Pliocene climate change. Shortly before the capture of the drainage of the Snake River through Hells Canyon, there was a significant change in the depositional environment in the Snake River Plain. This change may have been caused by a climate change.

Boise Valley Starts to Form

After deposition of the Tuana and Tenmile Gravels, the Boise River, as well as all rivers entering the western Plain, began an incision or canyon cutting phase indicating a lowering of base level. The Tenmile terrace was developed shortly after the beginning of the Pleistocene ice age; and since we are still within this ice age, all the terraces were developed during the Pleistocene and were directly affected by it. Because the drainage basin of the Boise River covers a vast glaciated area in the Sawtooth and Smoky Mountains, flows of the Boise River during periods of glacial melting may have been more than 10 times greater than historical flows. These high river flows, in response to changing climates during the Pleistocene, combined with the downcutting episodes of the Boise River, made the terrace development possible. So, each terrace is characterized by downcutting succeeded by deposition of coarse gravels over the newly formed terrace. The terrace deposits are similar from terrace to terrace and consist of poorly sorted pebble to cobble gravel in a matrix of coarse sand. Grain size decreases downstream and the sedimentary structures are indicative of high-energy braided-stream deposits.

Bruneau Sand Dunes

When wind loses its velocity and its ability to transport the sand it has carried from the surface, it deposits it on the ground. Sand deposits tend to assume recognizable shapes. Wind forms sand grains into mounds and ridges called dunes, ranging from a few feet to hundreds of feet in height. Some dunes migrate slowly in the direction of the wind. A sand dune acts as a barrier to the wind by creating a wind shadow. This disruption of the flow of air may cause the continued deposition of sand. A cross section or profile of a dune in the direction of blowing wind shows a gentle slope facing the wind and a steep slope to the leeward. A wind shadow exists in front of the leeward slope which causes the wind velocity to decrease. The wind blows the sand grains up the gentle slope and deposits them on the steep leeward slope.

Bruneau Sand Dunes

The Bruneau Sand Dunes State Park, established in 1970, is located about 8 miles east-northeast of Bruneau and about 18 miles south of Mountain Home. Camping facilities are available and a lake on the north side of the large sand dune may be used for boating and fishing.

Although there are many small dunes in the area, two large, light-gray overlapping sand dunes

cover approximately 600 acres. These two imposing dunes are striking, particularly because they dwarf most of the nearby land features. The westernmost dune is reported to be one of the largest (not the largest) single sand dunes in North America, standing about 470 feet above the level of the lake.



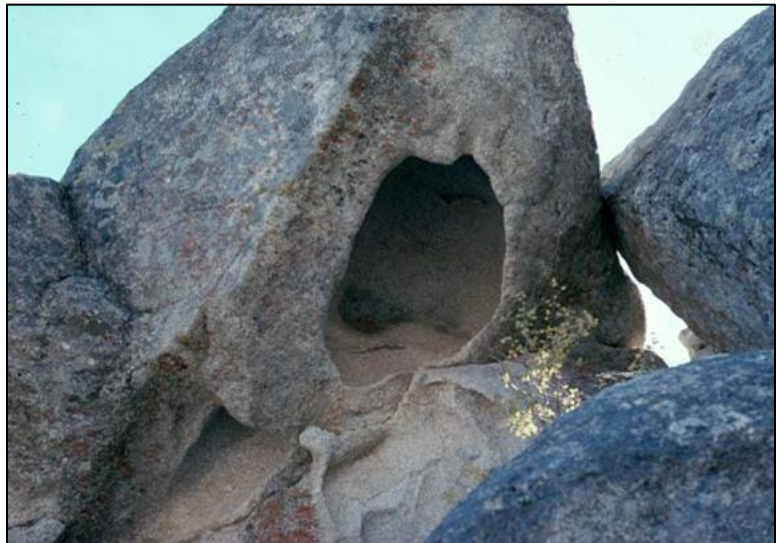
Large sand dune at Bruneau Sand Dunes State Park.

The existence of these dunes is attributed to the constant wind blowing sand in from the southwest. As the wind loses velocity over the basin, sand is deposited on the dunes. Sand has been collecting in the basin for more than 30,000 years.

Silent City of Rocks

City of Rocks National Reserve

The "Silent City of Rocks" which has been designated as the "City of Rocks National Reserve." covers a square mile area in Cassia County, approximately 4 miles from the Idaho-Utah border. It is situated 15 miles southeast of Oakley and about 4 miles west of Almo. You can reach the City of Rocks by traveling through Oakley on the west or through Almo on the east; both routes involve travel on graded gravel roads.



Hollow boulder from cavernous weathering of granite at the Silent City of rocks in south-central Idaho. Opening is about 3 feet in diameter. Penetration of the hardased shell of the boulder allows the interior to be removed by erosion.

Geologic Setting

The City of Rocks is situated in the Cassia Batholith. This small batholith covers more than 60 square miles in the southern part of the Albion Range. The batholith was at one time covered by a thin shell of Precambrian quartzite. Once the upper shell of protective quartzite was eroded away, the granite below eroded down at a more rapid rate. Consequently the City of Rocks is situated in a basin. Within the City of Rocks, more than 5,000 feet of granitic rocks are exposed from the top of Cache Peak to the bottom of the basin.

Pegmatite Dikes

Scattered pegmatite dikes, which have the composition and texture of coarse-grained granite, may be observed throughout the City of Rocks. Pegmatite ranges from thin seams to lenticular bodies up to 50 feet across and several hundred feet long. One exceptionally large pegmatite crops out in the City of Rocks. This pegmatite may be one of the largest to be found in Idaho with exposed dimensions of 200 to 300 feet wide and 400 to 500 feet long. Large masses of orthoclase feldspar, quartz and muscovite are well exposed over two rounded knolls that expose the pegmatite. Some of the masses of quartz and feldspar are tens of feet in diameter. Masses of muscovite display radiating crystals. Smoky quartz and miarolitic cavities are common. Numerous small workings over this large pegmatite show evidence of past interest and activity.

Weathering

Although Jointing controls the general form of outcrops in the City of Rocks, weathering is the agent responsible for creating the bizarre and fantastic shapes that characterize the area. On the surface of the outcrops, weathering occurs by granular disintegration. In other words, one layer of crystals after another is successively removed from the surface. This leaves the newly exposed surface in a smooth and rounded condition with no sharp or ragged edges or corners. The detrital material weathers from the granite and is carried by wind and water to low areas among the prominent forms. The grains of quartz, feldspar and mica at the surface of outcrops are friable and easily disintegrated with hand tools.

Chemical weathering occurs by solutions which penetrate the cleavage cracks in crystals and between mineral grains. Once the solutions are in these narrow boundaries, they cause new minerals to form which have a larger volume than the space available. This process of hydration and other chemical changes cause the disintegration and exfoliation.

Case Hardening

In addition to granular disintegration, case hardening is important in developing the unusual erosional forms. In some areas, an outer layer has been hardened by the deposition of other minerals such as iron oxides. Once a form has a case-hardened protective shell, the granular surface material is removed much more quickly underneath the shell. In some cases only the protective outer shell is left. In this way, caves, niches, arches, bath tubs or pans, toadstools and hollow boulders are formed. The case-hardened crust is generally darker in color than the lighter under sides.

Jointing

Jointing is exceptionally well developed in the Almo pluton and is particularly important in controlling the basic rock forms in the City of Rocks. There are three basic joint sets: one has subvertical dip with a northwest trend, another has a subvertical dip with a northeast trend, and a third has a horizontal trend and may represent unloading joints. Jointing controls the arrangement

of the outcrops and facilitates the weathering process by providing a plumbing system for solutions to migrate into the outcrops and cause the alteration, hydration and disintegration of the surface layers of granite. At some outcrops, weathering has caused joint spaces to widen to the extent that blocks are separated and form tall, isolated monoliths such as spires and turrets.

Twin Sisters

The Twin Sisters are both composed of granite, but are of vastly different ages. The tall spire is the 29-million-year-old Tertiary “sister” and the spire to the south of it is the 2.5-billion-year-old Archean “sister.” The Tertiary granite is composed of plagioclase, biotite and quartz and is intruded by muscovite-bearing aplite dikes. The gneissic granite of the Archean sister has similar composition but is megacrystic (relatively large crystals), is strongly foliated (parallel layers of biotite mica), weathers brownish, and is cut by dikes of the Tertiary granite.

The Gooding City of Rocks

The Gooding City of Rocks is situated in the Mt. Bennett Hills about ten miles northwest of the city of Gooding, Idaho. The Bennett Hills form an east- west-trending range of rounded hills which rise some 1600 feet above the surrounding Snake River Plain. Elevations in the area range from 4400 feet at the south edge to 6200 feet on the north. Drainages in the area include Fourmile Creek on the east, and to the west, Coyote, Dry, Cottonwood, and Clover Creeks.



Looking north at rock forms at the Gooding City of Rocks. North of Gooding, Idaho.

Structural Setting

The Mt. Bennett Hills are located on the northern edge of the Snake River Plain in an area where the Cenozoic volcanic rocks overlap the Idaho Batholith. The hills consist of a horst (an uplifted block of the Earth's crust) bounded by the Camas Prairie graben (a depression) on the north and the Snake River Plain downwarp to the south. They are made up of a sequence of volcanic rocks of Miocene age which have been intruded by rhyolites of Pliocene age. The hills rise abruptly out of the Snake River Plain along east-west-trending faults and then dip gently south to plunge beneath the Snake River Plain basalts.

Landforms Made of Tuff

The major landforms making up the City of Rocks are highly-dissected plateaus with deeply

eroded and deeply cut stream channels. The many streams and their tributaries though many flow only in the spring of the year, have helped dissect the volcanic rocks into weird and exotic forms. These spectacular landforms occur primarily in a geologic formation known as the City of Rocks Tuff, a member of the larger Idavada Group. A tuff is a fine-grained rock formed mainly of glass particles in which crystals of feldspar, quartz and other minerals are imbedded. The deposits are believed to have been produced by the eruption of dense clouds of glowing volcanic glass in a semi-molten state. As the material fell to the ground the glass particles were welded and fused together. Many layers are solely composed of glass and obsidian (volcanic glass) formed within the welded tuff formation.



Unusual rock form at the Gooding City of Rocks.

Joints Control Landforms and Features

The many pillars and arches in this tuff are a result of two forces; structural deformation and mechanical erosion. Weathering processes such as freezing and thawing have sculpted the rocks along multidirectional joint patterns (natural fractures in the rocks) to create a myriad of exotic features. For instance, closely-spaced horizontal joints make some columns appear like stacks of coins, whereas converging horizontal and vertical joints create weak points which are typically weathered to form concave features. Creation of the concave features such as undercut pillars, mushroom caps, and arches are also products of the weathering process. Alternating hard and soft layers in the rocks allow differential weathering and erosion to further shape the surface.

In some locations within the City of Rocks, major northwest-southeast-trending joints can be seen to form "fins" of rock oriented in the direction of those joints. The fins and other geologic features in the City of Rocks are quite similar to those found in the Arches National Monument in Utah.

Hoodoos

Some of the more unusual landforms found within the area are columns, arches, monoliths, and especially hoodoos. A hoodoo is a rock formation that originates in the following way: rocks are fractured during periods of faulting; water then penetrates the cracks and causes them to widen. This leaves a pillar of rock separated from the parent rock. The alternating hard and soft layers of the tuff give an undulating or wavy appearance to the hoodoo. These features are common in

volcanic areas and are usually concentrated in regions where most of the rain fall occurs during a short period of the year.

Big Wood River

A stretch of the bed of the Big Wood has world-class examples of bedrock erosional features. This remarkable geologic attraction is situated approximately 10 to 13 miles northeast of the town of Shoshone and is about 3 miles east of Kinzie Butte.

Basaltic Bedrock in the Big Wood River Channel

The basalt flow that hosts the Big Wood River channel has been dated by Armstrong et al (1975) near Magic Dam as 0.8 million years old. This flow displays extensive weathering and most of the swales and depressions are filled with loess which tends to smooth out the rough flow-top surface. The color of the basalt, which is formed of compound pahoehoe flows, ranges from light gray to very dark gray or black. In places where the channel has eroded through this basalt to a depth of more than 50 feet, at least three individual flows or cooling units are exposed along the canyon walls. At the upper few feet of the flows, features such as lava blisters, columnar joints, ropy pahoehoe, vesicular zones, and flow foliation are common.



Two potholes will ultimately enlarge and coalesce to form an elongated pothole. Big Wood River, Idaho.

Big Wood River after the Last Glacial Age

The Big Wood River originates from a number of tributaries draining out of the mountains to the north. From the reference point of Ketchum, Idaho, the Big Wood River has major tributaries running off the Boulder Mountains to the northeast, the Pioneer Mountains to the east and the Smoky Mountains to the West. Most of the pebbles and cobbles in the Big Wood River in the vicinity of the study area were derived from these mountains. This upland area was dramatically shaped by glaciers during the last period of alpine glaciation which ended around 7,000 to 10,000 years ago. When the present position of the Big Wood River channel was established approximately 10,000 years ago, there were enormous flows of glacial melt water running off the high country. These large flows carried heavy sediment loads and were instrumental in developing the spectacular potholes during the first few thousand years of the Big Wood River's existence in its present position.

Pebbles and Cobbles

One of the remarkable features of the channel is the abundance of diverse pebble and cobble types derived from the glaciated mountains more than 30 miles to the north. Almost all of these pebbles and cobbles are harder or more durable than the host basalt. Where pebbles and cobbles were trapped in a depression, the current drove the loose rocks around in a circular pattern grinding out the hole or recessed area into a bowl-shaped feature or pothole. As the pebbles skipped and bounced along the bottom by the current, their impact caused the pothole to deepen. As the potholes enlarged, they became cylindrical in form with smooth polished walls. In this manner, potholes up to 40 feet deep and 22 feet in diameter were created by the current-driven pebbles.

A surprisingly large number of rock types and compositions are represented by the pebbles. For the most part, they are much harder and more resistant than the host basaltic rock, therefore making them more effective as sculpting tools. The most common rock types of the pebbles include igneous intrusive rock, quartzite and gneiss. Except for the basalt, all pebbles are derived from drainages in the Pioneer Mountains, the Boulder Mountains and the Smoky Mountains north of the Snake River Plains. These pebbles and cobbles are too large to have been transported such a great distance under present flow conditions of the Big Wood River. They represent further evidence that the Big Wood River carried substantially more water in the past than it does at present.

Pothole Development and Evolution

Most of the basic erosional forms along the Big Wood River channel are created by numerous and diverse potholes in all stages of development and in a variety of shapes and sizes. The most common shape for a pothole is that of a teardrop--small at the top and large at the base. In fact the very largest potholes have a teardrop shape. Some potholes pinch and swell, some are cone shaped and some are cylindrical in form. At the base or lowest part of the pothole, most are bowl shaped.



Woman sitting in a teardrop shaped pothole that has merged with the very large pothole from which the photograph is taken. Notice the large-scale horizontal ribbing on the large pothole approximately where the woman's knees are positioned. When the large pothole merged with the teardrop-shaped pothole, the pebbles (grinding tools) of the teardrop-shaped pothole were captured through the hole by the larger and deeper pothole. Big Wood River, Idaho.

As you walk through the deeper part of the channel or canyon you can readily see many examples of "suspended" or "hanging" potholes at all levels along the smooth walls of the large potholes that are responsible for the present size of the canyon. These suspended or hanging potholes are generally exposed as a vertical section of less than one half of the original pothole. Existing in almost every size and shape, the suspended or hanging potholes are the relics or remnants of captured or cannibalized potholes. On the walls of some of the larger potholes, one can identify as many as 20 hanging, relict potholes in every stage of development. By studying these features on the canyon wall, it is possible to work out the sequence of development or evolution of the deep channel.

Development of Asymmetry

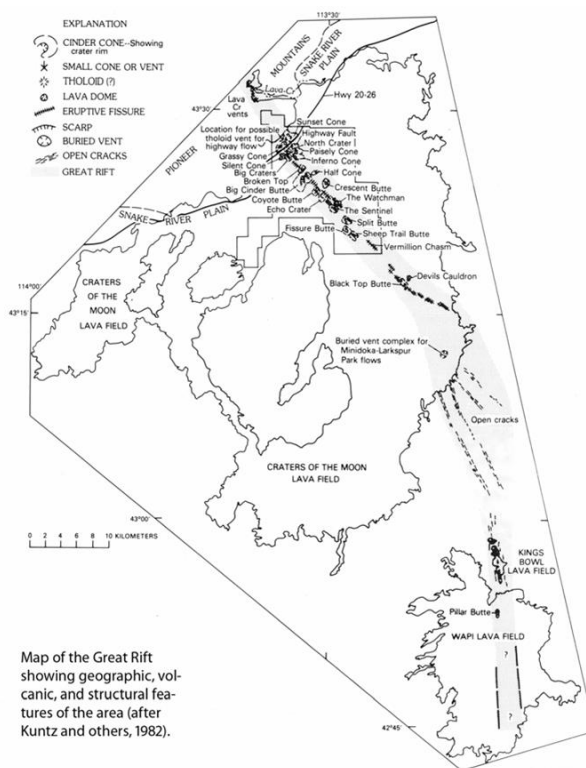
All of the features within the Big Wood River Channel are overprinted with a strong asymmetry. This asymmetry is imposed by the strong current-driven pebbles and cobbles against the upstream side of the features. Consequently, the upstream side of features tends to be smooth, convex and rounded; whereas, the downstream side tends to be concave with the leading edge of the feature pointing in the downstream direction. When looking at the channel in a perpendicular direction to the course of the channel, one can easily see the asymmetric development; similar to asymmetric ripple marks, they have gentle, convex slopes on the upstream side and relatively steep, concave slopes on the downstream side. Almost all the shapes in the channel were created by the erosional processes of asymmetry development, overprinting the features resulting from pothole development.

Age of the Big Wood River Channel

When the Black Butte lava flowed westward over the 0.8 million year old lava some 10,000 years ago, the Big Wood River was rerouted, possibly as much as one mile to its present channel. Therefore, the sculptured rocks of the Big Wood River are approximately the same age as the Black Butte basalt. Because the newly crystallized Black Butte lava is up to 10 feet higher than the underlying older basalt, the Big Wood River now follows the irregular flow front of the Black Butte flow.

The Great Rift

The Great Rift system consists of a series of north-northwest-trending fractures, which



extend 50 miles from the northern margin of the eastern Snake River Plain, southward to the Snake River. In 1968, the Great Rift was designated as a national landmark. The system has been divided into four separate sets of fractures. These four sets from north to south include: (1) the Great Rift set and cuts across the Craters of the Moon National Monument; (2) the Open Crack rift set and apparently has not experienced extrusive activity; the King's Bowl rift set; and (4) the Wapi rift set. The total rift system is 62 miles long and may be the longest known rift zone in the conterminous United States.

Recent Volcanism of the Eastern Snake River Plain

Very fresh basalt can be found at five different locations: the Cerro Grande and other flows near Big Southern Butte; Hells Half Acre lava field near Blackfoot; Wapi lava field; Craters of the Moon lava field; and King's Bowl lava field. The last three originate from the great rift system. The younger flows lack vegetation so that they clearly stand out on aerial photographs.



Aerial view along the northeast trending Great Rift. The rift can be followed about 60 miles across the central Snake River Plain. Basalt was extruded from the fracture and flowed both east and west creating a mirror-image effect.

King's Bowl Rift Set

The King's Bowl Rift set includes a central fissure with sets of symmetrical tension cracks on both sides. It is about 6.5 miles long, 0.75 mile wide and trends N. 10 degrees W. The main fissure is about 6 to 8 feet wide and is mostly filled with breccia and feeder dikes. In certain areas it is possible to descend into the rift several hundred feet. King's Bowl, which was created by one or more phreatic eruptions, is the most prominent feature on the rift. The ash blocks and rubble around the crater are evidence for an explosive eruption. Well-developed pipe-shaped vesicles are exposed in a massive flow on the east wall of King's Bowl. These vertically-aligned vesicles indicate the path of gas escaping from the base of the flow. The direction of the flow is indicated by the bend in the vesicles.

Split Butte

Split Butte, situated about 6 miles southwest of King's Bowl, is believed to be a maar crater. The name refers to a split or gap in the upper tephra layers at the east side of the butte. Prevailing west winds have caused the tephra ring to be asymmetrical. The winds caused more pyroclastic debris to be piled on the east side. The split, which is located on the east side, is believed to be caused by wind erosion. The tephra ring had an explosive pyroclastic phase. When the first flow

erupted, it passed through ground water. This caused glassy ash to form due to the cold water coming in contact with the hot lava. After the water saturated sediments were sealed, pyroclastic activity ceased and a lava lake formed. The lava lake partly overflowed and then crusted over. After withdrawal of liquid lava below the crust, the central portions of the crust collapsed.

King's Bowl

King's Bowl is a crater 280 feet long, 100 feet across and 100 feet deep. It stands directly over the main fracture of the Great Rift. Kings Bowl crater is the source of the 2,222 year old King's Bowl lava flow. Immediately west of the crater is an ejecta field where large blocks of rubble blown from the vent are strewn all over the ground. The size of blocks decreases with distance from the crater. A field of squeeze-up structures nearby was caused by lava being squeezed up through fractures. Some are hollow indicating that lava was drained out shortly after formation. The ash and ejecta fields were caused by ground water coming in contact with lava upwelling from the vent.

Wapi Lava Field

The Wapi Lava Field is located at the southern end of the Great Rift System. The Wapi lava field is a broad shield volcano covering approximately 260 square miles. The cone consists of many aa and pahoehoe flows which are replete with lava tubes and channels. The Wapi lava field formed about 2,270 years ago, almost simultaneously with the King's Bowl lava field.

Sand Butte

Sand Butte is located 23 miles southwest of Craters of the Moon National Monument. Sand Butte first formed as a tuff cone and was later filled by a lava lake. It is situated on a 3-mile-long, north-south-trending fissure. The fissure ranges from 200 to 410 feet wide. Sand Butte, like Split Butte, was formed by the phreatomagmatic interaction of ground water and basaltic magma. Pyroclastic flow was the primary method of deposition.

The basalt was erupted after the phreatic phase ended. This is demonstrated by tongues of spatter which overlie the tephra deposits. The final event was the partial filling of the crater by a lava lake. Finally the lava lake subsided to form a shallow crater.

Craters of the Moon

Craters of the Moon represents one of the most exceptional geological wonders in Idaho. It is a veritable outdoor museum of volcanic features. While traveling through the area between Carey and Arco,



Spatter cones aligned along the Great Rift. Craters of the Moon National Monument, Idaho.

the land darkened by basalt flows may, at first glance, appear to be a monotonous landscape. However, once you enter Craters of the Moon National Monument and inspect the area at close range, you will find a great variety of volcanic features. The Loop Road and the network of trails are designed to give you a self-guided tour of one of the great geological museums of the world. Features such as crater wall fragments, cinder cones, mini-shield volcanoes, lava bombs, spatter cones, lava tubes, tree molds and rifts are among the best examples, both in variety and accessibility you can observe anywhere in North America. The Craters of the Moon lava field covers 7 cubic miles of lava flows and pyroclastic deposits, including more than 60 lava flows and 25 cinder cones.

The Great Rift is a 53-mile-long and 1 to 5-milewide belt of shield volcanoes, cinder cones, lava flows and fissures. Three lava fields are aligned along the rift: Craters of the Moon, Wapi, and Kings Bowl. Craters of the Moon lava field covers an area of 640 square miles and is the largest Holocene (less than 10,000 years old) lava field in the conterminous United States.

In the area of the Great Rift, the Snake River Plain is broad and flat. It has lava flows at the surface with thin lens-shaped beds of eolian sand and alluvial sand deposits about 3,300 to 6,800 feet thick. The rift zone is aligned at right angles to the long axis of the eastern Snake River Plain.

Ages of the Lava Fields

Rift area was dated by charcoal from tree molds and beneath lava flows. This charcoal gave radio-carbon ages of 2,222 years old, plus or minus 60 years, for the King's Bowl lava field. Paleomagnetic measurements have also been used to correlate lava flows and determine the volcanic history in the area.

The flows were extruded during eight eruptive periods beginning about 15,000 years ago until the last eruption about 2100 years ago. Each eruptive period lasted several hundred years with a separation between eruptions of several hundred to several thousand years. Another eruptive period is likely to occur within the next 1,000 years because the last eruption occurred about 2,100 years ago.

As more and more flows are dated in the Snake River Plain, it is possible to relate the radiometric date to the surface character of a flow. For example, as a flow ages, the surface becomes lighter, more deeply weathered, more oxidized, more dust covered and more vegetated.

Eruptions along the Great Rift

The Great Rift is the source of lava that formed the Craters of the Moon, King's Bowl, and Wapi lava fields. Along the 1 to 5-mile-wide rift you will find a belt of cinder cones and eruptive fissures. The magma, which formed the flows and other features you now see at the surface,

originated from chambers of magma at depths of 40 to 90 miles. This magma, being less dense than the surrounding rock, rose along the deep fracture system of the Great Rift.

Excellent examples of eruptive fissures exist in craters such as the Trench Mortar Flat area between Big Cinder Butte and the Watchman cinder cones. The first eruptions along the fissures were characterized by fountains of lava, commonly referred to as “curtains of fire.” These curtains of fire were active along a fissure for a distance of up to two miles. In addition to the large flows that were extruded from the fissures, ridges and spatter ramparts were built along the fissure. In the final stage of activity along the fissure, spatter and cinder cones were formed.

Lava Flows

As each new lava flow was extruded at Craters of the Moon, it followed the lowest elevation, just as water would. Consequently, a number of high areas or knolls called “kipukas” are completely surrounded by lava flows. Several kipukas can be observed at Craters of the Moon. Geologists classify the fine-grained, black rock at the monument as basalt. Basalt, the most abundant igneous rock, consists of microscopic mineral grains (mostly plagioclase feldspar), volcanic glass and gas vesicles (air bubbles). Basaltic lava is erupted at temperatures of about 1800 degrees and flows over the ground surface as large tongues of molten rock.

At Craters of the Moon, you can see excellent examples of the three types of basaltic lavas: blocky, aa and pahoehoe. Blocky flows consist of a surface of smooth-faced blocks. Aa flows are blocky or platy but the fragments are covered with spines and have a very rough surface. Pahoehoe flows are characterized by thin sheets with low flow fronts. At the surface they are ropy, smooth or filamented. The upper surface of the fresh pahoehoe is made of a dense to vesicular glass. This surface is commonly a greenish blue or an iridescent blue. The Blue Dragon flows, which make up most of the surface east of the Great Rift, have an unusual iridescent-blue color. In a single flow, one lava type can change to another; for example, pahoehoe lava can change into aa lava when it cools, becomes more viscous and loses gas.



Typical pahoehoe flow with smooth, ropy, filamented surface at Craters of the Moon National Monument, Idaho.

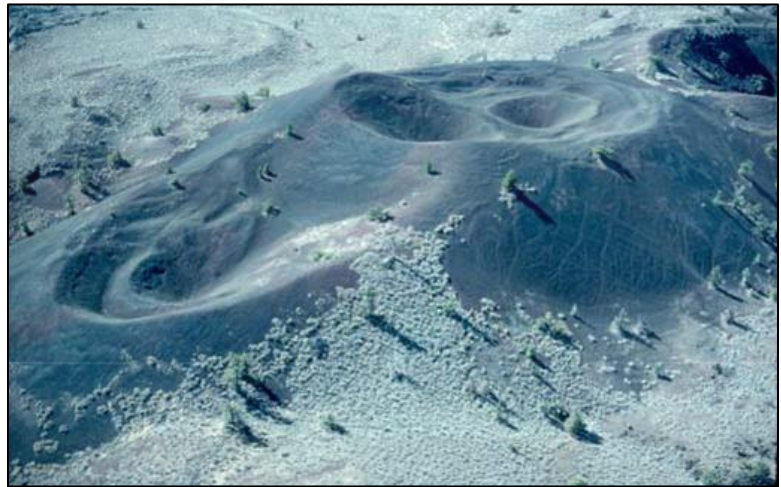
Most of the flows in the Craters of the Moon consist of pahoehoe-type lava and were fed through a system of lava tubes. In many locations, the roofs of lava tubes have collapsed, leaving a mass

of broken lava and rubble on the floor of the tube. Many such collapsed areas in the Broken Top and Blue Dragon flows provide “skylights” and entrances to lava tunnels for the numerous visitors to the monument.

The older flows in Craters of the Moon lava field can be identified by the nature of the surface. Younger flows have unweathered glassy crusts and a surface that is typically blue. Older lavas tend to be covered by wind-blown deposits and their surface is light colored from weathering and is strongly oxidized. A few of the flows in the craters area have an aa type surface. Aa flows have a rough, jagged surface.

Cinder Cones

Numerous cinder cones exist throughout the Snake River Plain. However there are more than 25 exceptionally well developed cinder cones aligned along a 17-mile-long, northwest-trending segment of the Great Rift. Cinders are very vesicular clots of lava blown out of a vent in the Earth’s surface. Cinder cones are developed by the accumulation of cinder and ash in cone-shaped hills. Many cinder cones at the monument are a composite of two



Cinder cone with multiple craters at Craters of the Moon National Monument.

or more cones with overlapping craters and flanks. These cones consist of agglutinated and nonagglutinated ash layers typically interlayered with a few thin lava flows. Some cones, such as the Inferno Cone and the Paisley Cone are asymmetric or elongate reflecting the wind direction at the time of the eruption. As a general rule, winds from the west caused more downwind accumulation of cinders on the east sides of cones.

Volcanic Bombs and Spatter Cones

Volcanic bombs are blown from volcanic vents as clots of fluid lava. As they move through the air while still hot and plastic, they deform into aerodynamically-shaped projectiles of lava. Spatter cones are relatively small cones formed by blobs of molten rock hurled out of volcanic vents. The hot, plastic blobs weld to the outer surface of the cone and quickly harden into rock.

Hagerman Fauna

In 1988 the Hagerman Fossil Beds National Monument was established by Congress. In 1928 Elmer Cook, a local rancher discovered fossil bones and showed them to H.T. Stearns, a U.S. Geological Survey Geologist. Stearns forwarded the bones to J.W. Gidley at the U.S. National Museum. Gidley interpreted the bones to be a fossil horse and during the next two summers he removed tons of bones. The horse, identified as *Equus simplicidens*, is the earliest recognized example of the modern horse genus *Equus*. Genetically *Equus simplicadens* may be more zebra than horse. The fossil bones were represented by specimens of all different ages and genders. This concentration of bones may have occurred when a herd of horses crossing a river were swept down river by a strong current and deposited in an inside meander bar where they were rapidly buried.



Overview of Hagerman horse quarry with town of Hagerman in the background. Note the bedding of the Pliocene Lake Idaho deposits.

In 1988 the Idaho State Legislature designated the Hagerman horse the State fossil. The Glens Ferry sediments in the Hagerman area have yielded the world's greatest variety and numbers of animals from the late Pliocene. These vertebrates include 18 fish, 4 amphibians, 9 reptiles, 27 birds and 50 mammals. More than 100 vertebrate species together with fresh-water clams and snails and plant pollen allow accurate interpretation of the nature of the landscape and ecosystem at the time. The Hagerman fauna perished some 3.5 Ma when the landscape was a broad, flat flood plain on the southeast edge of ancestral Lake Idaho.

During the Pliocene, the Snake River Plain received about 20 inches of water per year so vegetation was more diverse and susceptible to willows, alder, birch and elm as well as small concentrations of pine. Plant pollen provides evidence of plant species and their associated water requirements. Fossils of pelicans, cormorants, grebes, herons, egrets, storks, swans, ducks, geese and rails require an aquatic environment; also, fish bones, frogs, turtles, muskrats, beaver and otter are further evidence of water-dependent fauna.

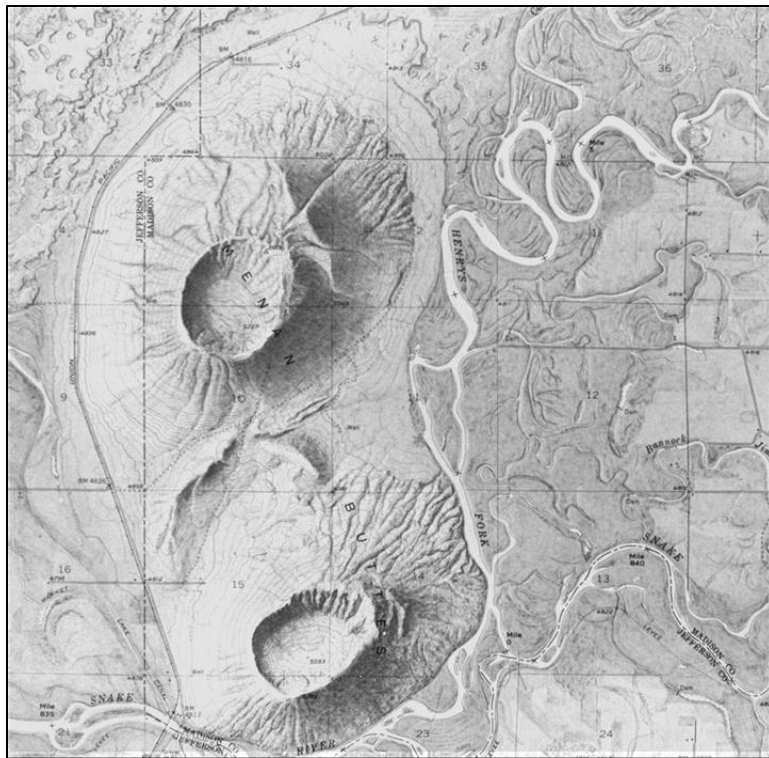
The presence of species such as horse, camel, peccary and antelope require a habitat of grassy plains with nearby water. The flood plain environment offers the necessary periods of flooding to sweep up and concentrate the various species and the associated rapid sedimentation as the flood abates to bury them.

Menan Buttes

The Menan Buttes are located on the Snake River Plain about 20 miles north of Idaho Falls in southeastern Idaho. Five cones were erupted along a north-northwest-trending fissure. Both of the two large cones are approximately two miles long in a northeast direction and about two-thirds as wide. Each cone has a large summit crater several hundred feet deep and about one-half mile long. The northern cone stands about 800 feet above the surrounding plain and the southern cone about 550 feet.

Eruption Diverted the River

The two cones consist of glassy olivine-basalt tuff. They were erupted through the water-saturated flood plain of the Snake River. As a result of the buildup of the tuff during the eruption, Henry's Fork of the Snake River was diverted to the south so that it now joins the Main Fork of U.S. Geological Survey map showing Menan Buttes, southeast Idaho. Note how the buttes formed in the flood plain of the Henrys Fork River and caused the river to be diverted. Also note the asymmetry of the cones caused by wind blowing pyroclastic materials toward the northeast. the Snake River at the east edge of the south cone.



U.S. Geological Survey map showing Menan Buttes, southeast Idaho. Note how the buttes formed in the flood plain of the Henrys Fork River and caused the river to be diverted. Also note the asymmetry of the cones caused by wind blowing pyroclastic materials toward the northeast.

Shape of the Cone

Most of the volcanic vent eruptions in this part of the Snake River Plain are basaltic shield volcanoes. Steep-sided cones built of tuff are rare. The asymmetric shape of the cones was caused by prevailing wind blowing towards the northeast. This wind carried most of the pyroclastic fragments blown out of the crater towards the northeast. Therefore the southwest slopes are steeper than the northeast slopes.

Composition of the Cones

The cones were built of glass fragments blown explosively from the crater vent. The successive

tuff beds from each eruption are conformable so that the beds inside the crater dip towards the center. Those beds on the outer slopes of the cones dip outwards at angles of 20 to 30 degrees. The layers range in thickness from less than an inch to several feet thick. The tuff is dark olive gray in color and consists almost entirely of glass.

Cone Contains River Gravels

Because the cones were erupted through the riverbed, they contain many xenolithic (foreign rock) cobbles, pebbles and sand grains from the river bed. The largest xenolithic pebbles and cobbles are most prevalent near the crater's edge. Some rounded. cobbles are as large as one foot across. This rounded river-derived material consists of aplite, granite, quartzite, basalt and rhyolite. Approximately 1 to 2 percent of the material in the cone was derived from the river gravels.

Steam-Charged Eruptions

The unique character of the tuff was caused by fluid lava making contact with the water in the river alluvium. Upon contact with water, the lava chilled and was erupted explosively by steam. Small fragments of lava were probably solidified to glass while driven explosively by steam from the crater.

Below the cones, the river gravels are interbedded with basalt flows. Blocks of this basalt, up to 5 feet in diameter are numerous near the crater rim and give one a good idea of the power of the steam-charged eruptions.

Age of the Buttes

Both buttes are approximately the same age which is believed to be very Late Pleistocene to Early Recent. Eruption of the cones caused a southwest diversion of the Henry's Fork. Before the eruption, the Henry's Fork flowed where North Butte now is situated.

Hot Springs of Idaho

Hot water bubbling out of the ground has been known about and used in Idaho for thousands of years. Concentrations of artifacts at and petroglyphs on rocks near hot springs are evidence that the areas were known about and used by Indians. More than 200 hot or warm (thermal) springs occur in Idaho and are distributed throughout several geologic provinces. In order for a spring to be classified as thermal it must be at least 68 degrees F. Some of Idaho's hot springs reaches temperatures as high as 200 degrees F. Even higher temperatures have been encountered in deep wells.

Origin of Hot Springs

Hot springs develop when rain and melted snow waters infiltrate into the ground. This ground water then sinks deep into the Earth and is warmed by the heat contained in the Earth's interior. Because hot water is less dense it is pushed back to the surface by the continued sinking of heavier, cold water. Annual replenishment to the system by cold meteoric water creates a continuous system which is called hydrothermal convection. Hot springs develop when the upwelling hot water rises along faults or other fractures in the Earth's crust and flows out onto the surface.

Sources of Heat

Why do some areas of the Earth have hot springs and others do not? For example, there are few if any hot springs identified in Idaho north of Idaho County. While there is heat in the Earth beneath any spot on the surface, this heat is concentrated closer to the surface in some areas. One means of near-surface concentration is a shallow magma chamber within the Earth's crust such as would exist beneath a volcano. Yellowstone National Park is centered on such a volcano. We know from measuring the way in which the vibrations of distant Earthquakes pass through the Yellowstone area that there is a body of molten rock (magma chamber) at a depth of about 3 to 6 miles. The resulting hydrothermal systems are world famous.

Another area where the heat of the Earth can be concentrated near enough to the surface to generate hydrothermal systems is where the crust of the Earth is stretched and thinner than normal. Such a condition exists over southern Idaho in the Snake River Plain and Great Basin areas. A number of large warm water systems occur in this region of Idaho including the Boise, Twin Falls, Bruneau-Grandview, Raft River, and Mountain Home systems.

Hot to warm springs also occur within the rocks of the Idaho Batholith. The heat in the batholith is believed to result from the decay of radioactive elements contained in many of the minerals which commonly occur in the granitic rocks.

Relationship of Faults to Hot Springs

Now that we have our sources of heat, we need a path for getting the hot water to the surface. In an active volcanic area such as Yellowstone, faulting is created by upward pressure exerted by rising magma, by explosive eruptions, and by deflation of a depleted magma chamber after an eruption. The latter mechanism is the most common and forms what is called a ring fracture system which approximates the outline of the magma chamber. Yellowstone has the added advantage of being astride a zone of active north-south to northwest-trending faults known as the Intermountain Seismic Belt. The broken rocks within the fault zones in Yellowstone act as an excellent conduit for circulating ground-water.

The Snake River Plain and Great Basin area of southern Idaho are being stretched. This is a very

advantageous geologic situation for the formation of hydrothermal systems because this not only creates a thin crust to bring heat closer to the surface but it also develops numerous normal faults and keeps them relatively open so that water can easily circulate through them.

Hot springs in the batholith are also the result of deep circulation of ground water in fault zones. It appears from recent geologic mapping that some of the normal faults of the Great Basin continue into the batholith. The batholith has also been subjected to other episodes of faulting.

Mineral Deposits

An interesting feature of hot springs is their associated mineral deposits. Yellowstone, for example, contains several types of hot springs deposits, some of which have resulted in spectacular rock formations.

Minerals are more readily dissolved in hot water. The type of minerals deposited depends on the temperature of the water and the type of rocks it passes through on its way to the surface. Mammoth Hot Springs in Yellowstone is a large hot spring deposit of calcium carbonate (calcite). Table Rock near Boise is a hot spring deposit composed of silica, mostly in the form of chalcedony. At Table Rock, near Boise, the hot water passes through mostly volcanic rocks and some sedimentary rocks that have been derived from the granitic rocks of the Idaho Batholith. These types of rocks generally contain about 60 to 70% silica. Once again the relationship is evident. Hot water dissolves the silica out of the rocks and redeposits it near the surface as the water cools and the silica drops out of solution.

Boise Geothermal System

To put our concepts to work and reinforce them let's take a look at an individual system. The Boise system is one of the most studied within Idaho and is generally representative of systems in the southern part of the state. This system is located on the northern margin of the western Snake River Plain graben. The heat source for this area of Idaho is the stretched and relatively thin crust. For reasons that are not well understood the heat in the western Snake River Plain is concentrated along the margins of the Plain. This is fortunate because the faults along which the plain is downdropped also occur along the margin. The main Boise Front fault can be readily seen as a distinct topographic break between the plain and mountains along the north edge of Boise. Comparison of rocks in the foothills of the Boise Front with those encountered in wells drilled in the plain has further confirmed the existence of offset along this fault and in fact has made possible the identification of a series of faults. The water of the Boise system circulates to depth along this series of faults, is heated, and is driven back to the surface by hydrothermal convection. In Boise, springs no longer flow at the surface because wells drilled into the fault zone to tap the hot water have intercepted and withdrawn the hot water rapidly enough that it no longer reaches the surface.

There is an interesting feature of the Boise system that may explain why it and other systems along the margin of the plain are located where they are. Recent geologic mapping has identified a major northeast-trending fracture system within the Idaho Batholith which appears to extend through the Boise geothermal area. This fracture system provides an excellent channel for the ample meteoric water which falls as rain and snow on the Boise Front to infiltrate into the ground and finally into the high temperature zones along the margin of the Snake River Plain. Further work may make it possible to correlate other fracture systems in the batholith to hot springs along the northern margin of the Snake River Plain and thus explain why hot springs do not occur at regularly-spaced intervals.

Borah Peak Earthquake

East-central Idaho awoke to a crisp autumn morning on Friday, October 28, 1983. Suddenly, the peaceful mountain morning was shattered as powerful forces within the Earth's crust were unleashed. Normal morning activities were interrupted as window panes vibrated and alarmed people ran out of their houses.



Looking north at the fault scarps along the western edge of the Lost River Range south of Dickey Peak.

The Earthquake occurred at 8:06 a.m. (MDT) with an epicenter located in a sparsely populated mountainous area between the small rural communities of Challis and Mackay. Challis and Mackay, Idaho lie in the northwest-trending valley on the west side of the of the Lost River Range. Updated calculations indicate that the Borah Peak Earthquake registered a Richter magnitude of 7.3. The epicenter was located about 19 miles northwest of Mackay, at the south margin of the Thousand Springs Valley just west of 12,662-foot Borah Peak. The Earthquake caused two deaths in Challis and an estimated \$15 million



The 1983 Borah Peak earthquake registered 7.3 on the Richter scale and resulted in a 35 km long fault scarp. Several subvertical normal faults with cumulative vertical displacement of 3 m disrupted the gravel road. Northwest of Mackay, Idaho.

damage in the affected region.

Geologic Setting

The Earthquake occurred along a fault zone on the southwestern flank of the Lost River Range. The core of the Lost River Range is composed mostly of folded and thrust-faulted Ordovician through Pennsylvanian sedimentary rocks including limestone, dolomite, quartzite, siltstone, and sandstone.

The Lost River Range is one of several northwest-trending mountain ranges in east-central Idaho. The topography is typical of the Basin and Range Province. Ranges are separated by broad sediment-filled valleys, and have range-front faults on their southwest flanks.

Ground Shaking

Primary effects of the Borah Peak Earthquake included ground shaking and surface rupture directly related to fault movement along the western flank of the Lost River Range. Ground shaking was most intense near the epicenter between Challis and Mackay, but the Earthquake was also felt over most of the northwestern United States and in parts of Canada. The initial shock was followed by numerous aftershocks. In the 10 months following the main Earthquake, there were at least 20 aftershocks with a magnitude of 4.0 or greater.

Secondary effects produced as a consequence of the Earthquake included seismically induced landslides, ground cracking, and modification of the hydrologic system. Damage to roads, buildings, and other structures occurred in the area between Mackay and Challis.

Ground Rupture Produced by Fault Displacement

Fault displacement that produced this intense Earthquake was expressed in spectacular surface ground rupture along a northwest-trending, 22-mile-long zone on the western flank of the Lost River Range. A west-northwest-trending section of faulting which branches off of the main surface fault trace west of Dickey Peak gives the surface faulting pattern a Y-shape. Much of the zone of surface rupture follows the Holocene and upper Pleistocene fault scarps of the Lost River fault. Fault scarps, the most common features along the zone of surface faulting, look like small steps or cliffs. Fault scarps are produced when adjacent blocks of the Earth's crust move relative to each other and are displaced along a fault plane.

Detailed study of the surface faulting and focal mechanisms reveals that the dominant fault movement associated with the Borah Peak Earthquake was dip slip, or vertical. That is, the Lost River Range was uplifted vertically along the fault relative to the adjacent valley. Maximum throw (vertical displacement) measured along the west flank of Borah Peak is about 9 feet. A more minor component of left slip is also evident, indicating that the Lost River Range block also moved laterally northward, relative to the valley to the west, as well as upward. The high

relief and linear northward-trending mountain ranges in this region and elsewhere in the Basin and Range Province have been produced by similar repeated fault movements over geologic time.

The surface fault trace is complex and comprised of multiple, gently curved, subparallel fault scarps extending along a north-northwest trend. The zone of extensive ground rupture is more than 300 feet wide in some locations. The complex nature the fault rupture can be clearly observed in the area immediately north of Birch Springs. Along this section of the mountain front, two to four en echelon (subparallel) scarps are visible. The maximum scarp height north of Birch Springs is as much as 16 feet. The newly created fault scarps dip from 70 degrees to near vertical and face westward toward the valley. However, within a few days after the Earthquake, many fault scarps, particularly those in well-sorted stream sand and gravel deposits, were rapidly degraded to angle of repose dips of about 30 degrees. At the base of the mountain-front slope, a small down-dropped block or graben approximately 75 to 100 feet also formed.

Landslides Induced by Seismic Shaking

Landslides, rock falls, rock slides, and other ground failures were induced in an area of approximately 1600 square miles surrounding the epicenter as an immediate result of the intense seismic shaking. The steep and rugged terrain of the Lost River Range, and of the adjacent Salmon River, Boulder, and White Knob Mountains significantly contributed to the susceptibility to landsliding as a result of Earthquake shaking in the region. Earthquake-triggered slope failures occurred in a variety of materials including colluvium, glacial deposits, talus slopes, and in fractured bedrock areas.

The majority of the landslides were rock falls or rock slides. Particularly susceptible to rock falls and rock slides were steep and rocky slopes composed of Challis Volcanics. Open joints and fractures, typical of weathered Challis Volcanics, produced inherent slope instability in this unit allowing large blocks to be easily loosened with seismic shaking.

Seventy-five miles north of the epicenter along steep roadcuts and cliffs on Highway 93 near Salmon, the road was severely obstructed by rock falls and rock slides originating in outcrops of Challis Volcanics. Dozens of boulders of Challis Volcanics, up to 10 feet in diameter, tumbled down steep slopes of Challis damaging several houses. Some boulders rolled as much as 200 feet out into the valley.

Largest Earthquake in 24 Years

The Borah Peak Earthquake was the largest Earthquake to hit the western United States in 24 years and was one of the only six historic Earthquakes of magnitude 7.0 or greater in the Basin and Range recorded since 1872. The Earthquake had significant immediate effects on the land surface, hydrologic system, and man-made structures. The Borah Peak Earthquake remains as a

reminder of the dynamic nature of the Earth's processes.

The Stanley Basin

The Sawtooth National Recreation Area covers about 314 square miles and includes parts of the Challis, Sawtooth and Boise National Forests. The area is accessible by many roads and trails. The trails are normally usable by foot from July to October, enabling hikers to reach the high country. Although the trail system is excellent, travel on foot away from trails is extremely difficult because of the rugged terrain.



Looking north-northwest into the Stanley Basin. The mountain range on the west side of the basin is the Sawtooth Range. This portion of the range was significantly modified by alpine glaciation during Pleistocene ice age.

The Sawtooth Range was named because of its jagged sawtooth profile along the skyline. Relief in the area ranges from an elevation of 5,000 feet where the South Fork of the Payette River leaves the area to 10,830 feet at Thompson Peak - the highest peak in the range.

In addition to the magnificent scenic beauty of the area, a large number of geologic features can be observed. The present land forms have been produced primarily by a combination of faulting, jointing and glacial ice. Most of the area making up the Sawtooth Mountains is underlain by rocks of the granitic Idaho Batholith and the younger Sawtooth Batholith.

Stanley Basin

The basin is a north-trending half graben that is part of the basin-and-range extension faulting of the last 17 million years. The seismically active Sawtooth fault marks the boundary of the west side of the Stanley Basin. There is no fault on the east side of the Stanley Basin. Throughout most of the Pleistocene all of the ranges were occupied by glaciers including the Sawtooth ranges on the west, the



Looking west across Stanley Basin at Red Fish Lake. The Sawtooth Mountains are in the background. These mountains have outstanding examples of all the classic erosional glacial features. The tree covered lower hills on the viewer's side of the lake are underlain by glacial lateral and terminal moraines which serve to dam the lake.

White Clouds to the northeast and the Boulder Mountains to the southeast and the Salmon River Mountains to the north. These mountains have outstanding examples of alpine glaciation.

The Sawtooth glaciers converged in the Stanley Basin and covered much of the basin floor. Glaciers from the White Cloud Mountains descended to the basin floor but did not coalesce. On the west side of the basin, a number of glacial lakes are impounded by moraines.

Faults

The Sawtooth Valley or Stanley Basin is a half graben (down-dropped block) bounded on the west side of the valley by the Sawtooth Fault. The basin also trends northwesterly from its origination at Galena Summit near the headwaters of the Salmon River. The basin is covered by glacial moraine and river deposits. The moraines are covered by lodge pole forests, and by contrast, the outwash flats are covered by sage brush.

The spectacular Sawtooth Range is a northwest-trending uplifted fault block, commonly called a horst. The range is bounded on the west side by the Montezuma fault and on the east by the Sawtooth fault. These northwest-trending faults are extensions of basin and range faulting north of the eastern Snake River Plain. Both faults terminate against the major Eocene Trans-Challis Fault System. The horst is composed of Cretaceous granodiorite, Tertiary (Eocene) pink granite and metamorphic roof pendants of possible Precambrian age.

The White Cloud Peaks, which form the range on the east side of the valley, are white calcsilicates formed by hydrothermal alteration of the Pennsylvanian-Permian Wood River and Permian Grand Prize Formations by the White Cloud stock. The stock has been dated at 85 Ma.

Glacial Features

Almost all features of mountain glaciation, both erosional and depositional, can readily be observed in the area. Many large, glaciated valleys with their characteristic U-shaped profiles dominate the area. Along these valleys, many hanging valleys or hanging tributaries may be observed. Cirque basins, matterhorn peaks, aretes and cols are examples of well-developed glacial features caused by ice erosion. Other evidence of moving ice includes polished and striated rock outcrops.

The topography is continuously modified by such processes as frost wedging, rock falls, talus and soil creep, rock-glacier flow and stream erosion. The high terrace behind Stanley is composed of glacial lake outwash.

Glacial Lakes

Most of the 500 plus lakes in the area were caused by glacial ice. In the high country, the lakes occupy shallow basins carved in the rocks; at lower elevations lakes tend to be contained by

lateral and terminal moraines. Examples of such impounded lakes include Alturas, Pettit, Yellow Belly, Stanley, and Redfish Lakes. Many of the cirques are occupied by cirque lakes.

Glacial Lake Stanley (not to be confused with Stanley Lake) existed in the Stanley Basin (near the town of Stanley) at a time when the Salmon River was dammed by glacial till. Excavations near Redfish Lake reveal a section of glacial fluvial (glacial river) sediments grading upward into a sequence of glacial lacustrine (glacial lakebed) sediments.

Glacial deposits consist of unsorted clay, sand, cobbles and boulders left by melting glaciers. This material was removed from the mountains and transported to the site of deposition by glaciers.

Ridges of Glacial Moraines

Along the eastern front of the Sawtooth Mountains is a series of elongated ridges standing approximately 1,500 feet above the Stanley Basin. These ridges may be distinguished from the bare, jagged Sawtooth Mountains in several ways. Because they are underlain by glacial till, they are rounded and covered by a thick stand of lodgepole pine. Redfish Lake occupies the lower end of a U-shaped valley and is contained by lateral moraines on the sides and a terminal moraine at the end near the lodge.

Rock Types

The Thompson Creek Formation is the oldest rock in the area and may be of Precambrian age. It consists of well-foliated mica schist interlayered with metamorphosed carbonate rocks. Where the carbonate rocks have intrusive contacts with the granitic rocks, skarn minerals may be well developed.

Light gray granitic rocks of the Idaho Batholith underlie about one-half of the Sawtooth Mountains. The batholith contains varying amounts of quartz, orthoclase, plagioclase and biotite, and has been dated at 108 million years.

The Sawtooth Batholith is a distinctive pink granite which contrasts sharply with the gray granitic rocks of the Idaho Batholith. The color is caused by the presence of perthitic orthoclase. The Sawtooth Batholith is younger than the Idaho Batholith. Dikes of the pink granite intrude the gray granite.

These Tertiary plutons were probably emplaced at less than 2 miles depth, whereas the batholith was emplaced at about 6 miles. The Tertiary plutons contain twice as much uranium and thorium than is found in the Mesozoic batholith. Many gas cavities contain smoky quartz, feldspar and aquamarine (beryl) crystals.

Joints

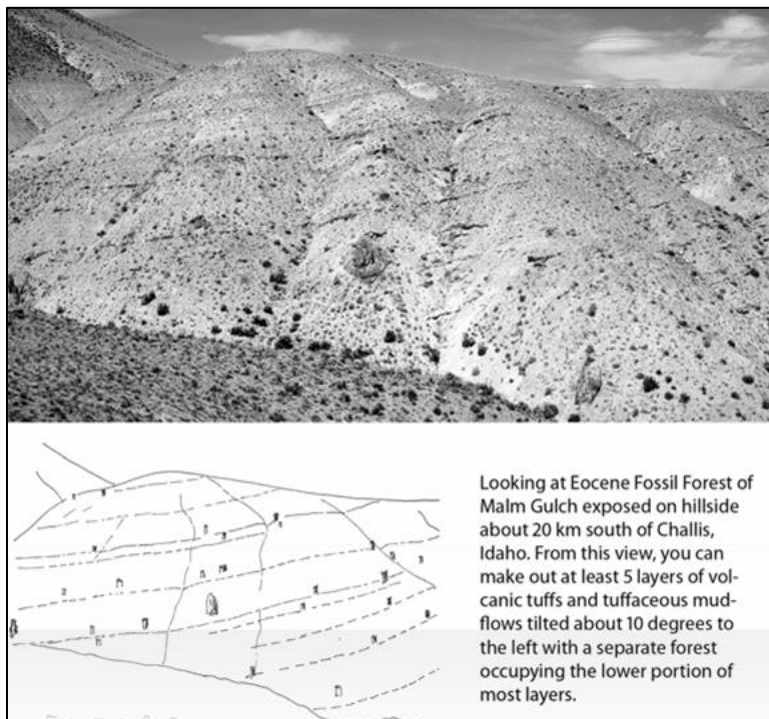
The granitic rock of the Sawtooth Range is thoroughly jointed. The joints dip steeply and are closely spaced. The two most common sets strike northeast and northwest. The jagged ridges of the range were developed because the joints made the rock susceptible to erosion by frost wedging.

Beryl (Aquamarine Deposits)

Beryl deposits have been known in the Sawtooth Mountains for many years. Beryl mineralization is generally confined to the Sawtooth Batholith or near its periphery. Areas of good beryl mineralization include the valleys of upper Pinchot Creek, upper Fall Creek, Benedict Creek, Spangle Lakes, Ardeth Lakes, Edna Lake and Toxaway Lake. Aquamarine occurs as scattered single grains, as northeast-trending zones of clustered grains and as sunbursts on joint surfaces. High-angle pegmatite dikes contain crystals of aquamarine in vugs or cavities. Aquamarine in pegmatites is associated only with quartz and orthoclase and is clear of inclusions. Aquamarine also occurs as spherical concretions of aquamarine, albite and quartz.

Malm Gulch

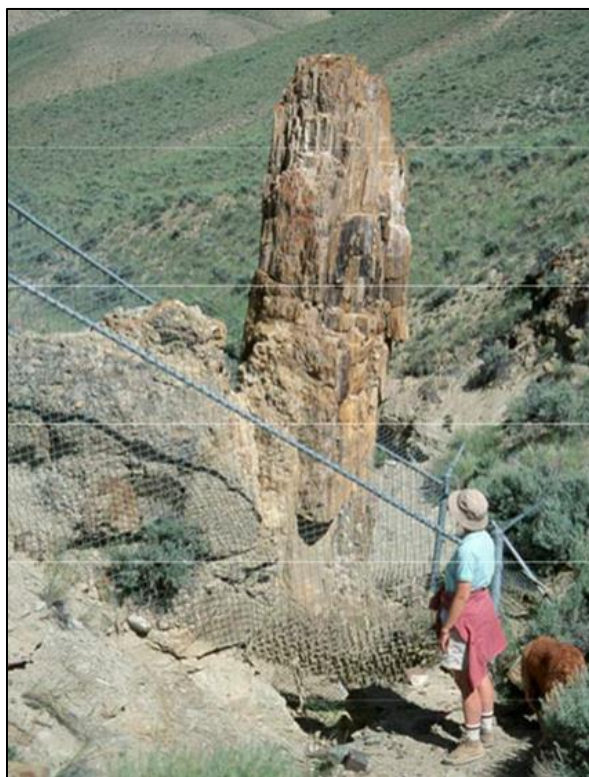
The petrified forest of Malm Gulch is in east-central Idaho about 12 miles south of the town of Challis. It is located about three miles from the east side of U.S. 93 up the gulch in a southwesterly direction. There are about 30 to 40 upright stumps of silicified Sequoia trees still in growth position and 10 to 15 horizontal logs exposed at the surface. According to Jones, there are at least six successive levels of forests distributed through a sequence of Eocene volcanic ash layers totaling about 175 feet thick. Each forest level represents a period long enough for soil to develop and trees to grow as much as 10 feet in diameter before being destroyed by air-fall tuffs and mudflows. A tree ring count indicates an approximate age of 500 years for a tree 9 feet in diameter. The average tree has a diameter of 6 to 8 feet at the base. Although most of the wood is silicified, it is not gem quality.



Looking at Eocene Fossil Forest of Malm Gulch exposed on hillside about 20 km south of Challis, Idaho. From this view, you can make out at least 5 layers of volcanic tuffs and tuffaceous mudflows tilted about 10 degrees to the left with a separate forest occupying the lower portion of most layers.

The BLM has placed wire cages around the stumps near the canyon bottom to prevent removal by collectors.

The petrified forest and surrounding environment has been designated as an “Area of Critical Environmental Concern” and a “Research Natural Area” by the Bureau of Land Management. Ross (1937) first collected botanical specimens in the Malm Gulch area. The Germer Basin flora consists of at least 37 species, including 2 horse tails, 2 ferns, 9 conifers and 20 angiosperms. This flora represents species that are typical of trees, shrubs and plants found in middle-latitude forests in a moist temperature climate like the present Pacific northwest. The petrified stumps of Sequoia trees (*Metasequoia occidentalis* and *Sequoia affinis*), some still in growth position, are exposed over several acres on the north side of Malm gulch. Each of the 7 forest levels represents a period between eruptions that was long enough for trees to grow as much as 10 feet in diameter before destruction by air falls of volcanic ash. The ash has been dated to be 48 to 50 million years old.



Large Sequoia tree still in growth position at Malm Gulch. The Bureau of Land Management has installed fencing to protect the resource of collector.

Plant remains include wood fragments, logs in horizontal position, stumps, fossil leaves, fruits and seeds. Some wood has been rounded by water, but many of the stumps are still in the position of growth. Most of the wood material has been silicified, but is not of gem quality.

Chapter Author:

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